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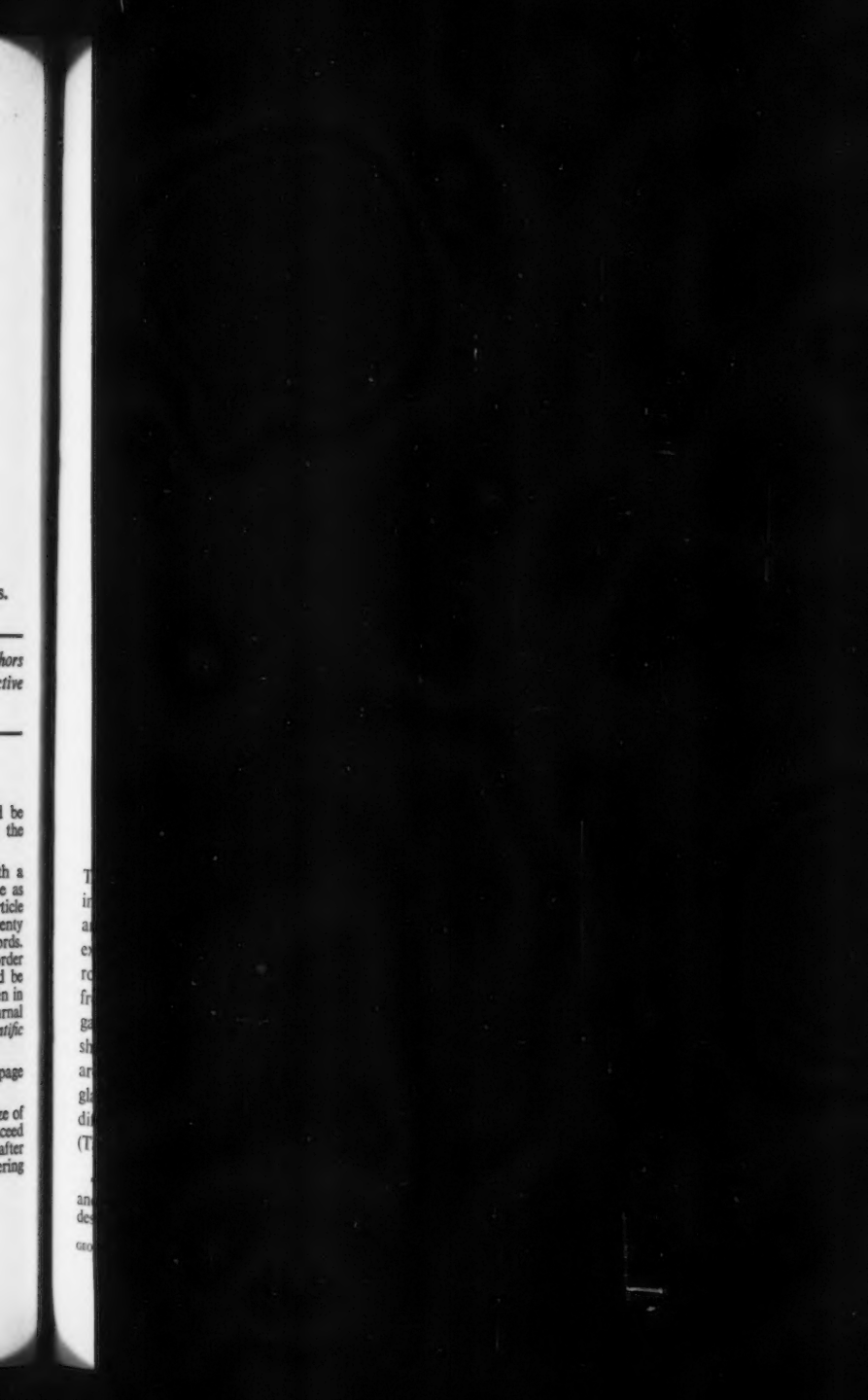


GEOLOGICAL MAGAZINE

VOL. XCVIII, No. 5. PUBLISHED 4.10.1961 SEPT.–OCT. 1961

A Note on the Origin of Ophitic Texture in the Chilled Olivine Gabbro of the Skaergaard Intrusion

By L. R. WAGER

(PLATES XVII–XIX)

ABSTRACT

The textures of the chilled marginal gabbro of the Skaergaard intrusion and of the coarser marginal gabbros 25 yards further in are described and related to counts of the number of separate crystals of the chief mineral species per cubic centimetre of the rock. An explanation of the textures and number of crystal individuals is attempted in terms of various degrees of supersaturation in the magma. In the chilled rock abundant nucleation of plagioclase and olivine, presumably at the middle labile stage of supersaturation, is inferred. This is followed by ophitic or poikilitic crystallization of pyroxene and ilmenite about scarce nuclei, produced at lower temperatures, presumably during the early labile stage of supersaturation for these two minerals. The coarser gabbro, further in from the margin, has about half the number of plagioclase crystals per cm.^{3} compared with the chilled rock and only one twenty-fifth the number of olivine crystals, and these have a tendency to be ophitic in texture. Under these slower cooling conditions, supersaturation apparently still reached into the middle labile region for plagioclase, giving many nuclei, while olivine, like pyroxene, separated under early labile supersaturation conditions and only relatively few nuclei formed. The contrasting textures of the plateau magma-type basalts of Mull and the tholeiitic basalts of Northern Ireland are explained along similar lines.

THE chilled olivine gabbro occurring at the margins of the Skaergaard intrusion contains pyroxene with an ophitic texture towards felspar and olivine, and this, in places, grades into a poikilitic texture *; an example collected 1 metre from the southern contact with the country rock is shown in the photo-micrograph, Plate XVII, A. A similar rock from the eastern margin is figured in the original paper on the Skaergaard intrusion (Wager and Deer, 1939, pl. 20, fig. 1). For the rock shown in the plate (EG 4507) the number of individual crystals per unit area has been estimated by marking 2 mm. squares in ink on the cover-glass of a thin section, counting the number of individual crystals of different species in a block of adjacent squares, and then averaging (Table 1). When counting, it is frequently necessary to use crossed nicols

* There is no general consensus of opinion on the usage of the terms ophitic and poikilitic but the meaning attached to them here will be clear from the description of Plates XVII–XIX.

to decide whether outlying patches of some of the crystals, especially pyroxenes, are in structural continuity. Even then the count is not precise; the observer has to carry in his mind the crystals already counted when he moves to another part of the square, and has to note whether a particular crystal projects into an already-counted square and if so, must be careful not to count it again. Ambiguities may arise due to a single twinned crystal being counted as two, or two adjacent crystals, having by chance an essentially parallel orientation, being taken as a twin and counted as one. In practice it has been found best to make two fairly rough counts rather than one very careful one—the first count being of crystals that are certainly distinct individuals (the minimum figures), and the second, of all that could possibly be regarded as distinct individuals (the maximum figures). The mean of these counts is considered to provide an adequate approximation to the number of individual crystals. Even though no great precision is attained, the considerable differences in the relative numbers of individual crystals of different species are clear enough and are useful in elucidating the sequence of events during the solidification of the magma.

From its appearance in thin section the iron ore is also considered to be ophitic. It occurs in clusters of small units with interstitial habit separated by considerable distances from other similar clusters. Vincent (1960, p. 997) has shown that the opaque iron ore in the chilled marginal gabbro is ilmenite and that the clusters of units are in fact in optical continuity when seen in polished sections, and so are correctly described as ophitic or poikilitic. In estimating the number of individual crystals of opaque minerals for this paper thin sections and not polished sections were used. Each cluster of opaque grains, separated by considerable distances from neighbouring clusters, has been assumed a single ophitic crystal and counted as such.

From the counts of the numbers of individual crystals over an area of the section of the chilled rock, the number per cm.^{3} has been calculated (Table 1). The figures show that for every pyroxene * crystal there are over one-hundred times as many plagioclases, about fifty times as many olivines and about a third of an ilmenite crystal. From their general appearance, it is clear that broadly similar results would be obtained for the other examples of chilled gabbro at the margin of the intrusion, except in cases where the magma has apparently been contaminated with acid gneiss.

Another rock type occurring about 10–70 m. from the contact has been called the patchy pyroxene rock; it is much coarser than the chilled rock but carries similar ophitic to poikilitic pyroxenes (Plate XVIII).

* In the marginal rocks considered in this paper there are two different pyroxenes present but since they are texturally similar they are treated together to simplify the discussion.

The numbers of individual crystals per cm.³ in an example of this rock have also been estimated (bottom of Table 1). The total number of plagioclases per cm.³ is 2,700, which is about half the number found in the chilled marginal gabbro. There are about 160 times as many plagioclase crystals as pyroxenes, a proportion not very different from

TABLE 1.—NUMBERS OF INDIVIDUAL CRYSTALS OF THE CHIEF MINERALS IN TWO ROCKS OF THE MARGINAL BORDER GROUP OF THE SKAERGAARD INTRUSION

A. Chilled marginal gabbro EG 4507, 1 m. from contact, southern margin

Square Counted (size 4 × 4 mm.)	Felspars Min.*	Felspars Max.*	Pyroxenes Min.	Pyroxenes Max.	Olivines Min.	Olivines Max.	Ilmenites Min.	Ilmenites Max.
No. 1	50	58	2	3	39	41	—	—
,, 2	38	49	3	4	21	23	2	2
,, 3	36	40	2	3	10	13	—	—
,, 4	31	34	—	—	30	32	—	—
,, 5	47	58	1	1	43	47	2	2
,, 6	39	47	2	2	18	18	1	1
,, 7			2	2	22	25	2	4
,, 8			1	1	12	13	—	—
,, 9			—	—	24	25	—	—
,, 10			2	3	23	30	1	1
,, 11			3	3	13	15	—	—
,, 12			1	1	24	29	1	2
Average per square	40·2	47·7	1·6	1·9	23·4	25·8	0·8	1·0
Mean of max. and min. figures	44 0		1·8		24·6		0·9	
Nos. per cm³	280		11		150		6	
Nos. per cm.³	4,600		38		1,900		13	
Mode (Vol. %)	59		25		15		1·1	
Av. Vol. of individual crystals in cubic millimetre	·13		6·6		·08		·8	

B. Coarse patchy pyroxene rock EG 4442, 30 m. from the margin, Ivnarmiut, western margin

	Felspars	Pyroxenes	Olivines	Ilmenites †
Nos. per cm.³	2,700	16	75	22‡

* See page 354 for meaning of these figures.
† A little magnetite is also present.
‡ This figure might be lower if the count had been made on polished specimens which would allow the crystallographic orientation to be observed.

that found in the chilled rock ; but there are only five times as many olivines as pyroxenes, contrasting with fifty times as many in the chilled rock. The number of distinct ilmenite crystals is rather doubtful because the counts were not made on polished specimens, but it is of the same order as in the chilled rock.

From the number of crystals of the different mineral species in a unit volume of the Skaergaard chilled olivine gabbro it is evident that a relatively large number of crystal nuclei of felspar and olivine were

established during cooling, but only a few of pyroxene and ilmenite, and the texture of the rock is obviously controlled by these differences.

It is believed that the chief factor producing large numbers of crystals of olivine and felspar, and few of pyroxene and ilmenite, is the order of appearance of these phases under equilibrium conditions, combined with the crystallization behaviour of supercooled liquids. For the Skaergaard magma the sequence of appearance of the phases during cooling under equilibrium conditions is known in detail for the part of the cooling history covered by the observable layered series. The sequence in the earlier stages is less certain and will remain so unless drilling allows sampling of the layered series beneath the level now exposed. However, recent studies of the border group, combined with evidence from the lowest observable layered rocks, suggest that plagioclase and olivine separated first, followed later by augite and still later by pigeonite. Thus the lowest layered rocks are without cumulus pyroxenes and this is probably true for the bulk of the hidden layered series. On the other hand, evidence from the gabbro picrite shows that under some conditions a magnesium-rich augite and magnesium-rich orthopyroxene can separate (Brown, 1957). Perhaps both of these, but particularly orthopyroxene, may be a temporary cumulus phase in the hidden layered rocks although probably not in the earliest. The evidence for the late appearance of pyroxenes relative to plagioclase and olivine is not unequivocal, but, in developing the present discussion, will be assumed *. For the late appearance of magnetite and ilmenite as cumulus minerals there is clear evidence from the layered rocks of the lower zone (Wager and Deer, 1939, see also Wager, 1960, fig. 1, Vincent, 1960, p. 998).

The relative numbers of individual crystals of the various species in the chilled marginal gabbro, together with their textural relations (particularly the ophitic and poikilitic habit of the pyroxenes and ilmenites) suggest the following sequence of events during the cooling of the magma against country rocks:—

First, rapid cooling of the magma so that strong super-saturation occurred for plagioclase and olivine, the higher temperature phases yielded by this particular magma. Due to this strong super-saturation, crystal nuclei of these two phases formed in abundance and grew to a moderate size. The first rapid cooling was apparently insufficient to lower the temperature far enough for nuclei of pyroxene and iron ore to be formed. The latent heat released by the first-separating plagioclase and olivine crystals probably maintained the temperature for some time above the limit for the precipitation of pyroxene and ilmenite.

* Compare, however, new information in the addendum to this paper.

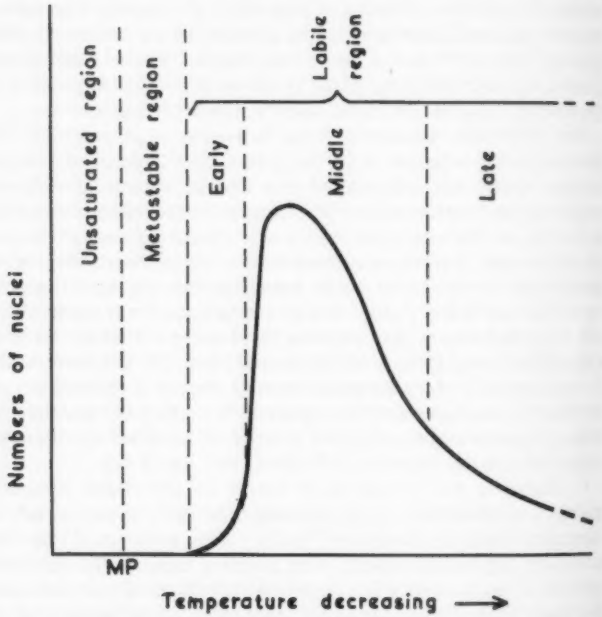
Second, after sufficient further cooling, the appropriate temperature conditions for the formation of pyroxene and, a little later, of ilmenite were reached. Only a few centres of crystallization of these minerals were established, and diffusion of material towards these allowed large crystals to grow. Since a large number of felspar and olivine crystals were already present, the pyroxene and ilmenite partially or completely enveloped them, giving the ophitic or poikilitic texture. The time needed for extensive diffusion of material to the centres of growth of the pyroxene and ilmenite would be provided by the necessarily slow cooling, due partly to the latent heat released by the precipitating plagioclases and olivine and partly to the steadily rising temperature of the country rocks as heat passed into them from the magma.

The conditions necessary for the formation of crystal nuclei in super-saturated solutions is by no means fully understood. Some workers accept the existence of two distinct regions of different crystallization behaviour, described long ago by Ostwald as metastable and labile. In the metastable region, crystallization does not occur at all unless seed crystals are introduced, while in the labile region spontaneous crystallization occurs, especially when the liquid is stirred or suffers mechanical shock. Some investigators, for example Miers and his collaborators, give evidence for a sharp distinction between the metastable and labile fields (Miers and Isaac, 1907). Others doubt the existence of a sharp boundary between the two contrasted sets of conditions ; they consider that there may be no metastable region, or only a very narrow one, and then a region of abundant spontaneous nucleation (see the summary in Buckley, 1951, pp. 7–42).

In discussing the crystallization history of the chilled marginal gabbro, a combination of the opposing views will be used which is diagrammatically summarized in Text-fig. 1. The existence of a narrow metastable region is accepted; then comes a region called here the *early labile* region where a few nuclei may spontaneously develop, and this passes gradually into the *middle labile* region where very abundant spontaneous crystallization occurs. Finally there is a *late labile* region where, as all workers agree, there is again very slow spontaneous nucleation. The arguments developed below to explain the textures of the chilled marginal gabbro would not be directly applicable if what is here called the early labile region did not exist, but it might well be a region which only becomes significant when cooling is slow.

The abundant nucleation of plagioclase and olivine relative to pyroxene and ilmenite, which produced the textural results seen in the rapidly chilled marginal rock, is ascribed to supersaturation reaching well into the region of labile conditions for plagioclase and olivine, giving abundant nucleation of these minerals. By the time the temperature was low enough for pyroxene to be separating under equilibrium

conditions, loss of heat would necessarily be slower, and the temperature would fall more gradually. Thus the conditions would be appropriate for the formation of a few nuclei of pyroxene in the early labile region and these would relieve the supersaturation, so that the middle labile region of abundant nucleation of pyroxene would not be reached. The slower cooling would allow enough time for diffusion to bring to



TEXT-FIG. 1.—Diagrammatic representation of the relationship between crystal nucleation and the degree of supersaturation as assumed for the purpose of this paper. The subdivision of the labile region into a so-called early, middle, and late stage is indicated.

the few scattered nuclei the required material for their growth, and supersaturation elsewhere in the liquid would thus be prevented from reaching into the middle labile region for pyroxene. Similar arguments may be used to explain the textural features of the ilmenite.

The relative numbers of crystals of different mineral species in the coarser rocks of the tranquil division of the marginal border group, such as EG 4442, 25 m. from the margin (Plate XVIII and Table 1), are significantly different from those of the marginal chilled rock, and so too

are the textures. Associated with these coarser rocks are certain special rock types, including the so-called perpendicular felspar rock, for a description of which see Wager and Deer, 1939, pp. 144–150. It has been realized for some time that the elongated felspars of the perpendicular rock must be the result of growth into a supercooled liquid, as with the olivines in the harrisites of Rhum (Wadsworth, in the press). The inner 90 per cent or so of the elongated plagioclase crystals is An_{70}. The outermost 10 per cent is reversibly zoned, and reaches An_{77}. The narrow zone of An_{77} is considered to be the solid material in equilibrium with the liquid, while the inner part of composition An_{70} apparently separated at a lower temperature from the supercooled liquid. Thus the perpendicular felspar rock provides evidence for considerable supersaturation during its formation.

The relative abundance of the crystals in the coarser rocks can also be interpreted in terms of supercooling. The abundant plagioclase individuals suggest supercooling until the middle labile region was reached for plagioclase. On the other hand, the much less abundant olivine crystals suggest that supercooling was not sufficient to reach into the middle labile region for this phase, but only into the early labile region where relatively few crystal nuclei of olivine were formed. This is believed to be due to the plagioclase being a mineral which, under equilibrium conditions appears at a somewhat higher temperature than olivine, in this particular magma.* This effect is not seen in the strongly supercooled chilled rock because the liquid was carried rapidly down to a temperature where both the plagioclase and olivine were in their respective middle labile nucleation regions. The labile shower of plagioclase crystals and their subsequent growth presumably slowed up cooling, and the slightly lower temperature required for olivine nucleation was only attained as a result of a more leisurely cooling which reaches only the early labile region. It is interesting to note that in these rocks, where the number of olivine nuclei was not very different from the number of pyroxene nuclei, the olivine crystals tend to grow into the spaces between the plagioclases and exhibit what may be described as sub-ophitic texture (Plate XVIII).

Fortunate circumstances of mineral composition indicate a greater degree of supersaturation in the chilled marginal gabbro than further in (cf. Brown, 1957, p. 530 and fig. 5). Thus, in some of the marginal border group rocks, 20 or 30 metres from the contact, nucleation of pyroxene occurred at a temperature just high enough for orthopyroxene and not pigeonite to separate directly from the magma, while in the chilled marginal gabbro, only a metre or so from the country rock, pigeonite, now inverted to orthopyroxene, was originally present. The

* This has been confirmed experimentally, see addendum.

reason for this is that the vigorous chilling close to the contact carried the temperature well down, before effective nucleation occurred, so that pigeonite separated. Under slower cooling conditions further in, effective nucleation occurred with less supercooling and a higher temperature, more magnesian, orthopyroxene was directly precipitated. The nature of the plagioclase indicates the same thing ; thus the innermost zone of the plagioclase in the strongly chilled rock is about An_{72} and, as mentioned above, it is believed that the first formed inner zones would have been An_{77} if the first crystallization had been under equilibrium conditions. Although, for the chilled Skaergaard gabbro, the conditions seem to have been fairly constant, with the production of a constant texture, further into the intrusion there was more variation, as shown by the occurrence of zones of perpendicular felspar rock and other special types. Crystal nucleation is highly sensitive to other factors besides the rate of cooling—such as currents in the liquid and the effect of already formed crystals which, although isomorphous, have sufficient structural similarity to increase the chances of new kinds of nucleus formation. The complexity of the phenomenon is certainly greater than pictured here, and a more detailed analysis must await a further description of the whole marginal border group of the Skaergaard intrusion.

Some Comparisons

Rough counts of the numbers of crystal individuals in two typical plateau magma-type basalts and two tholeiitic basalts have been made (Table 2). No great accuracy was attempted in the counting, the method of averaging a minimum and maximum figure being adopted ; the marked difference between the number of crystal individuals of different mineral species is, however, not in doubt.

For the plateau magma-type basalts, two rocks from Morven were chosen for measurement. The first, from Lochaline pier (Bailey *et al.*, 1924, Table I, No. II), has strongly ophitic pyroxene (Plate XVII, B) and there are about 700 times as many felspar individuals and 120 times as many olivines, as pyroxenes (Table 2). The counts for the other plateau magma-type basalt, also from Morven (Bailey *et al.*, 1924, Table I, No. III), shows broadly similar features. The difference in numbers of crystals is not considered to be the effect of the small differences in proportions of the different minerals, but to be due to the plagioclase, olivine and perhaps iron ores separating before pyroxene, under middle labile conditions, while pyroxene separated, as in the Skaergaard chilled rock, somewhat later and under early labile conditions.

For the two examples of tholeiitic basalts from Northern Ireland here considered, the relative numbers of the crystals of the various minerals are quite different : thus in the tholeiitic basalt of the Giant's

Causeway (Plate XVII, C) the number of pyroxene crystals is of the same order as the number of felspars and iron ores*, and in the overlying flow there are twice as many pyroxenes as plagioclases (Table 2). It is suggested that from these particular tholeiitic magmas, plagioclase, ore and pyroxene crystallized simultaneously from the start of crystallization. Supersaturation of all three is believed to have occurred and nucleation to have been at a middle labile stage.

These preliminary measurements cannot do more than suggest

TABLE 2.—NUMBER OF INDIVIDUAL CRYSTALS PER CM.^{3} OF THE CHIEF MINERAL SPECIES IN CERTAIN BRITISH TERTIARY BASALTS AND COMPARISONS WITH THE CHILLED MARGINAL GABBRO OF THE SKAERGAARD INTRUSION

	Felspars per cm.^{3}	Pyroxenes per cm.^{3}	Olivines per cm.^{3}	Iron ores per cm.^{3}
1. Plateau magma type basalt	350,000	500	60,000	40,000
2. Plateau magma type basalt	690,000	1,800	13,000	150,000
3. Chilled marginal Skaergaard gabbro	4,600	38	1,900	13
4. Tholeiitic basalt [1]	1,500,000	1,000,000	—	1,800,000
5. Tholeiitic basalt [1]	1,200,000	4,000,000	—	560,000

1. Plateau magma type basalt. Raised beach cliff, 200 yds. W. of Lochaline Pier, Morven (Mull Memoir, 1924, p. 15, No. II, Surv. Anal. 282).
2. Plateau magma type basalt. Lava (thick flow on ashy material). Little port, E. side of Rudha Dearg, 1 mile W. of Lochaline, Morven. (Mull Memoir, 1924, p. 15, No. III, Surv. Anal. 281).
3. Chilled marginal Skaergaard gabbro (see Table 1).
4. Tholeiitic basalt (Mid-lava Series). Grand Causeway, Giant's Causeway, Co. Antrim (Patterson, 1951, p. 286, Anal. No. 7).
5. Tholeiitic basalt (Mid-Lava Series) No. 2 flow. Quarry ½ mile N.N.E. Scudion Craig, Co. Antrim. (Patterson, 1951, p. 286, Anal. No. 8.)

[1] Counts exclude the crystallites in the glass.

possible explanations of the textures in terms of degrees of supersaturation. The contrasted habit of the pyroxenes in the two type of basalt here considered is usual. That the explanation of the ophitic habit of the pyroxenes in the plateau magma type basalt is due to the late appearance of this mineral is supported by observations by Elliott on the trachyophitic texture (1952, pp. 925–8). He rightly points out that this texture implies orientation by flow of platy plagioclase crystals before the growth of the poikilitic pyroxene, and he thus establishes the order of crystallization of these minerals in the cases he considered. In lavas, flow of the magma during crystallization must often affect the textures and in particular, when poikilitic pyroxenes were tending to form, movement must break them down

* The granularity of these rocks, given by Tomkeieff (1940, p. 108), is related to the figures given here for the number of individual crystals. In rocks with ophitic or poikilitic textures, however, granularity is not directly related to the number of individual crystals because various outlying parts of a single ophitic pyroxene crystal would be counted as distinct grains.

into numerous disorientated units. The absence of ophitic texture in some plateau magma type basalts may be due to this, but the characteristic absence of ophitic pyroxenes in the tholeiitic basalts has, no doubt, a more fundamental cause.

The relative numbers of crystals of different species at different distances from the margins of sills and dykes should provide useful evidence on the detailed processes of solidification. Winkler and his collaborators have made studies of the grain-size, which is inversely related to the number of crystal individuals if poikilitic and ophitic crystals are not involved, and have related the differences in grain size at different distances from the margin to the conditions of crystallization (Winkler, 1949, Winkler and Weitz, 1956).

In layered intrusions, where there is bottom accumulation of early crystals, a subophitic, ophitic, or poikilitic texture is usually shown by all minerals which are not cumulus phases, and there is some similarity in conditions of formation to the poikilitic minerals of the chilled Skaergaard gabbro. The Skaergaard intrusion, where the layered rocks are ortho- or meso-cumulates (Wager *et al.*, 1960), probably provides the simplest case. Slow cooling of the intercumulus liquid here resulted in only sporadic nuclei of the new crystal phases being produced, probably under early labile conditions, and diffusion of material to the sporadic nuclei allowed them to grow into poikilitic, ophitic, or subophitic crystals. As examples, cumulates from adjacent bands in the lower layered series of the Skaergaard intrusion may be cited. A plagioclase cumulate (Plate XIX, A) shows plagioclase crystals surrounded by intercumulus pyroxene, olivine, and iron ore with poikilitic habit ; in contrast, the adjacent olivine cumulate (Plate XIX, B) shows olivine crystals surrounded by poikilitic plagioclase, pyroxene and iron ore as intercumulus crystals. As with the chilled rock the poikilitic texture is related to late crystallization. If the late crystallizing mineral is fairly abundant, the texture will be poikilitic ; if relatively rare, the crystals will be small and have ophitic or subophitic relationships. The poikilitic, ophitic, and subophitic textures of the intercumulus minerals of cumulates are found to grade naturally one into the other and are the same phenomenon developed to a different extent.

The significance of the ophitic texture has often been discussed, e.g., Fenner (1926), Bowen (1928, pp. 67–9), Krokström (1932), Walker (1957), and there is as yet no consensus of opinion. In the chilled Skaergaard magma the relative number of crystal individuals which controls the texture, seems to be related to the differing degrees of supersaturation attained by the different minerals during the cooling process. The first-forming crystal phases, plagioclase and olivine suffer most supersaturation and nucleate abundantly in the middle labile region. The later crystal phases, pyroxene and ilmenite, for

under conditions of less rapid cooling ; thus nucleation takes place in the early labile region, and relatively few nuclei develop. At this stage the slower cooling allowed time for greater diffusion of ions and more extensive crystals grew. The textures are thus related to the order of crystallization combined with the effects of supersaturation and nucleation.

It should be noted that the ophitic texture of the pyroxene of the Skaergaard chilled gabbro does not mean that pyroxene was the only mineral crystallizing in the later stages as has sometimes been supposed. Using the evidence from the layered series for the sequence of crystallization, it is clear that plagioclase continued to precipitate right through the crystallization period. Thus, simultaneously with the growth of the extensive poikilitic or ophitic pyroxenes in the chilled marginal rock, there must have been steady addition of plagioclase material to the abundant crystals already present and this deposition must have continued, along with growth of pyroxene and ilmenite, to the end of the crystallization process.

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DEPARTMENT OF GEOLOGY
AND MINERALOGY,
OXFORD.

ADDENDUM, JULY, 1961

Dry melting experiments of the kind mentioned by Yoder and Tilley (1957) in the Annual Report of the Carnegie Institution for 1956–57 have kindly been carried out by Dr. H. S. Yoder and Dr. J. F. Schairer at the Geophysical Laboratory, Washington, on a sample of marginal chilled gabbro of the Skaergaard intrusion. The experiments are designed to determine the order in which mineral phases present in a powdered sample of a rock, melt as the temperature is raised, this order being regarded as the reverse of the order of crystallization during cooling under equilibrium conditions. In the case of the chilled marginal gabbro the experiments showed that plagioclase, pyroxene, and olivine are still present as crystal phases at a temperature of 1,200° C. At 1,215° C. plagioclase only is present. Plagioclase is also present, in small amounts, at 1,225° C. but by 1,250° C. it has disappeared and no crystalline phase remains. Yoder and Schairer suggest that the disappearance of the plagioclase, marking the liquidus, is likely to be about 1,230° C.

These experiments indicate that on cooling of the Skaergaard magma, under equilibrium conditions, plagioclase would be the first crystal phase to appear, then at a temperature about 20° lower, both olivine and pyroxene would appear approximately together.

In attempting to explain the texture of the Skaergaard chilled marginal gabbro it has been shown (p. 356) that plagioclase and olivine both nucleate abundantly before pyroxene, of which only relatively few centres formed. It was suggested that this might be connected with the fact that the plagioclase and olivine, under equilibrium conditions, appeared at a higher temperature than pyroxene, a view for which there was some indirect evidence (p. 356). It now appears that this sequence is wrong and that under equilibrium conditions plagioclase would appear first, at the highest temperature, and that pyroxene and olivine would appear later, approximately together, and at a lower temperature. Apparently some other explanation of the abundant nucleation of the olivine at about the same time as plagioclase is needed.

In the light of the experimental results it is now suggested that the

explanation of the textures of the Skaergaard chilled marginal gabbro should be modified to bring in the effects of differences in the ease of nucleation which are believed to be related to the differing structural complexity of the crystals (cf. Wager, 1959). Plagioclase probably nucleates with difficulty and so requires, in general, greater supersaturation to effect nucleation than olivine which probably nucleates easily, being a mineral of simple structure. Thus on rapid cooling of the Skaergaard liquid nucleation of any kind might be delayed until the temperature was, say, 25° below the liquidus. At this temperature we may suppose nucleation of plagioclase took place, the liquid having been supercooled 25° below the equilibrium temperature for plagioclase crystallization; at the same temperature the more easily nucleating olivine might nucleate abundantly, although the liquid was supercooled only by about 5° for olivine. Pyroxene may be expected to nucleate with more difficulty than olivine because of its more complex structure and thus its nucleation would occur at a lower temperature than the plagioclase and olivine and under what has been called early labile conditions.

The texture of the coarser rock, some 25 m. in from the margin and of essentially the same composition as the chilled marginal gabbro, shows that somewhat different conditions existed there during cooling. Thus plagioclase nucleated abundantly but olivine much less abundantly so that it is subophitic in texture and begins to resemble the habit of the pyroxene. In the case of rock further in from the margin the rate of cooling must be presumed to have been less; effective nucleation of plagioclase probably took place with less supercooling than in the chilled marginal rock, say at 15° below the liquidus temperature, and above the temperature for equilibrium separation of olivine. On further cooling olivine would separate but now under early labile conditions which would produce only relatively few centres of crystal growth and thus resemble texturally the pyroxene which probably separated not much later.

It is valuable to have this additional information on the likely sequence of crystallization of the initial Skaergaard liquid under equilibrium conditions and I am grateful to Drs. Yoder and Schairer for providing it.

EXPLANATION OF PLATES

PLATE XVII.—A. Chilled marginal gabbro of Skaergaard intrusion, EG 4507, 1 m. from the southern contact with the basalts. The number of individual crystals of plagioclase is 4,600 per cm.^{3} and of olivine 1900 per cm.^{3}. The number of ophitic and poikilitic crystals of pyroxene is 38 per cm.^{3} and of ilmenite 13 per cm.^{3}. × 9.
B and C. Contrasting textures of plateau magma-type basalt, Lochaline Pier, Morven (B) and tholeiitic basalt, Giant's Causeway

(C). B has 500 pyroxene crystals per cm.^{2} and 350,000 plagioclase crystals. C has 1,000,000 pyroxenes per cm.^{2} and 1,500,000 plagioclases. × 50.

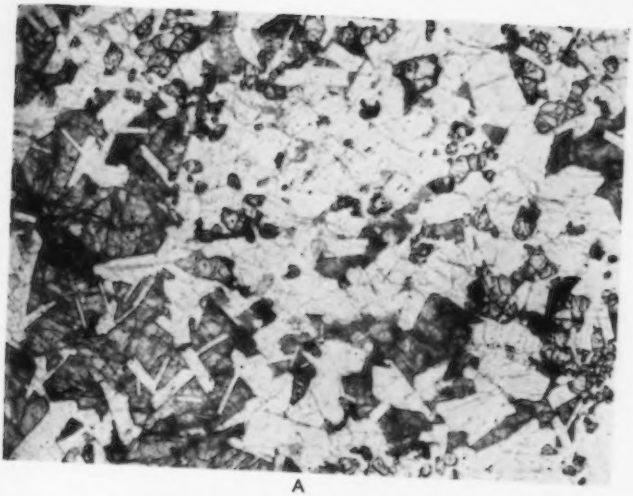
PLATE XVIII.—A. Coarse olivine gabbro, EG 4442, from the tranquil division of the marginal border group at 25 m. from the margin on Ivnarmiut, W. side of Skaergaard. The numbers of individual ophitic pyroxene crystals are 16 per cm.^{2} and of subophitic olivines 75 per cm.^{2} × 9.

B. Same rock as A but under partly crossed nicols showing the numerous individual plagioclase crystals amounting to 2,700 per cm.^{2}.

PLATE XIX.—Photomicrographs of contrasted neighbouring bands in the lower zone of the Skaergaard intrusion.

A. Plagioclase cumulate (EG 5109) with poikilitic pyroxene, olivine and iron ore as intercumulus material. × 9.

B. Olivine cumulate (EG 5108) with poikilitic felspar, pyroxene and iron ore as intercumulus material. × 9.



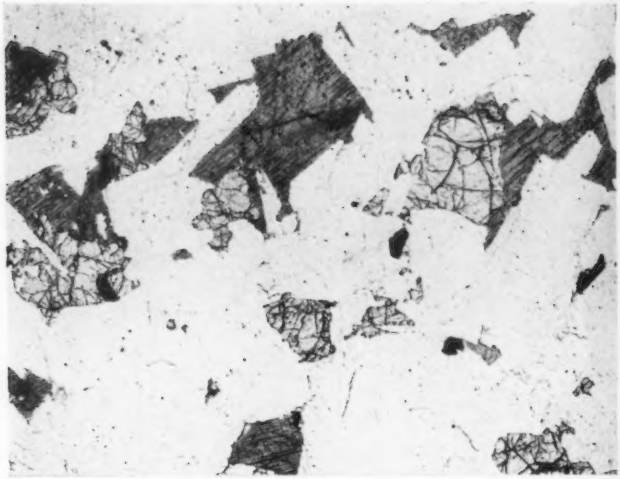
A

B

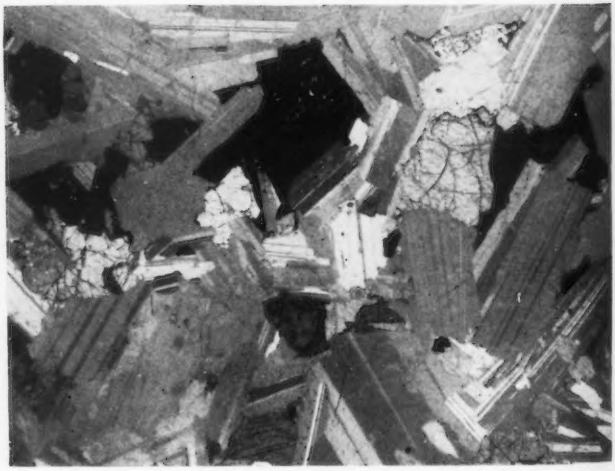
C

OPHITIC TEXTURE IN GABBRO.

facing p. 366.

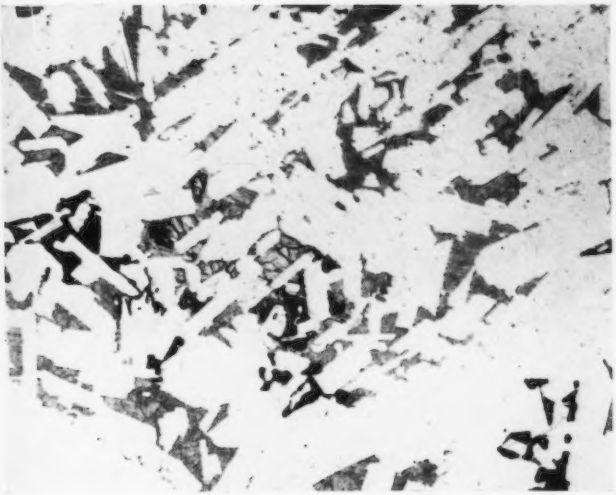


A

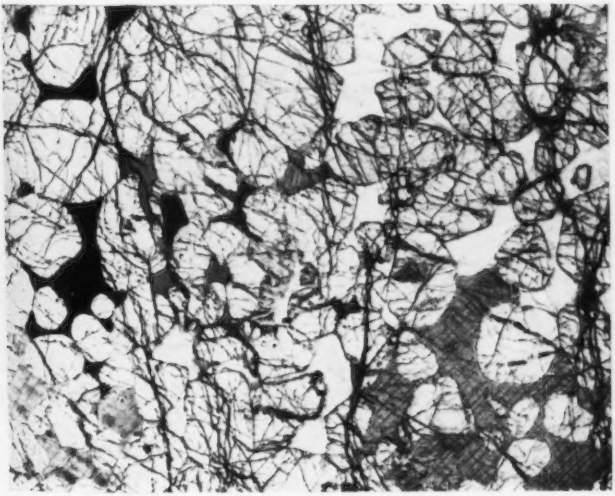


B

OPHITIC TEXTURE IN GABBRO.



A



B

OPHITIC TEXTURE IN GABBRO.

The Ordovician and Associated Pre-Cambrian Rocks of the Pontesford District, Shropshire

By W. T. Dean and D. L. Dineley

(PLATE XX)

Abstract

Following a recent re-mapping of the district east of Pontesford Hill, Shropshire, it now seems likely that the Uriconian and Longmyndian rocks at The Lyd Hole are faulted against one another, and that the Pontesford Shales form two outcrops, separated by faulting, and not a single outcrop as previously suggested. Doubt is cast on the validity of the so-called " rhyolite " of Habberley Brook, and the faunas and correlation of the Pontesford Shales are reviewed in the light of new evidence.

I. Introduction and Acknowledgments

The area of Shropshire described in this paper lies to the south and east of the village of Pontesford, between 6 and 7 miles south-west of Shrewsbury. In recent years it was mapped by geologists of the Geological Survey, the results being published by Whitehead (1929 ; *in* Pocock *et al.*, 1938). The district contains rocks of Uriconian, Western Longmyndian (Wentnorian) and Ordovician age, and lies at the northern end of the main outcrop of the Western Uriconian (see Text-fig. 1). In the vicinity of Habberley Brook the Pre-Cambrian rocks are overlain unconformably by Ordovician strata, mainly mudstones and shales, a tongue-shaped outcrop of which was shown on the Geological Survey's map to extend southwards from near Pontesford and along the eastern side of Pontesford Hill and Earl's Hill for approximately one mile. Exposures of solid rock are generally limited to steep banks in Radlith Wood and Oaks Wood, and to the course of Habberley Brook, particularly below Oaks Wood where the stream is deeply entrenched. The higher ground is, however, heavily mantled with boulder clay, and in the area between Earlsdale and Earlshill Farm augering has shown the superficial deposits to be over 7 feet thick. Yellow clay with fragments of Longmyndian (?) sandstone was found at the base of an auger hole 9 feet deep near Earlshill Farm.

The writers' attention was drawn to the area while conducting field-classes from Preston Montford Field Centre, and their results, some of which do not accord with those published previously, are reported here. The graptolites collected during the field-work have been identified by Dr. Isles Strachan, who has commented on their stratigraphical significance, and Professor W. F. Whittard has read and criticized the manuscript.

II. The Pre-Cambrian Rocks

It has been suggested by Whitehead (*in* Pocock *et al.*, 1938, pp. 33–5) that the dark and buff Uriconian rocks cropping out between the

Ordovician sediments of the north of the area and the " grits " above The Lyd Hole, south-east of Earlsdale Farm, reveal an ascending succession from west-north-west to east-south-east, and our observations confirm this. The junction between the Uriconian and Longmyndian rocks at the top of the waterfall at The Lyd Hole is fairly well exposed, and is seen to be remarkably straight and sharp. Although the disparity is small, the strike of the plane of this junction appears to be not quite parallel to the strike of the Longmyndian beds. Callaway (1882, p. 121) regarded the junction as a fault, but Whitehead, noting that " uncrushed rhyolite " and Longmyndian grit appeared to exist within a few inches of one another, rejected this view. Slickensides occur, not only within the rhyolite as stated by Whitehead, but are also well developed between the rhyolite and the sediments. The grit is locally shattered and eroded, and the junction in the side of the gorge at The Lyd Hole has every appearance of being faulted.

Since it is now generally accepted that the Western and Eastern Uriconian rocks are coeval, and that the Strettonian and Wentnorian subdivisions of the Longmyndian are successively younger formations, it would be expected that the junction of the Western Uriconian and the Wentnorian must be either a fault or an unconformity. No evidence of an unconformity has been found at The Lyd Hole, and a faulted relationship is suggested. The postulated fault-plane hades only a degree or two from the vertical and strikes approximately 12° east of north. There is nothing to suggest that it is other than a normal fault, possibly involving slow, gradual movement for no fault breccia has been found, though there is local quartz veining. It has not proved possible to map the northward extension of the fault.

III. The Pontesford Shales

(*a*) *General remarks*

The Pontesford Shales were described originally by Callaway (1877) who equated them with the Harnage Shales of the type Caradoc Series in the area east of The Longmynd. The geology of the Pontesford district and the stratigraphical position of the Pontesford Shales were discussed by Whitehead (1929 ; *in* Pocock *et al.*, 1938, pp. 90–95) who showed that the beds rest unconformably on Western Longmyndian rocks and that they contain abundant graptolites indicating the Zone of *Diplograptus multidens*. At the same time he claimed that the Caradoc Series east of the Longmynd exhibits a large stratigraphical break between the Hoar Edge Grits (Zone of *Nemagraptus gracilis*) and the Harnage Shales (said to belong to the Zone of *Dicranograptus clingani*), and on these grounds Callaway's earlier correlation was rejected. More recently a new interpretation of the type Caradoc succession has been suggested (Dean, 1958) whereby the Costonian Stage (Zone of

N. gracilis) is followed conformably by the Zone of *D. multidens*, which is thought to include both the Harnagian and Soudleyan Stages. On the basis of this interpretation any shelly fossils found in the Pontesford Shales should, then, belong to all or part of the Harnagian or Soudleyan Stages. After examining the small collection from Habberley Brook now in the Geolocial Survey and Museum it was suggested (Dean, 1958, p. 228) that the Glenburrell Beds (Lower Soudleyan Stage) of the Onny Valley were represented, and the present work has confirmed this suggestion. Graptolitic strata beneath those containing the shelly fossils are now equated with part of the Harnage Shales.

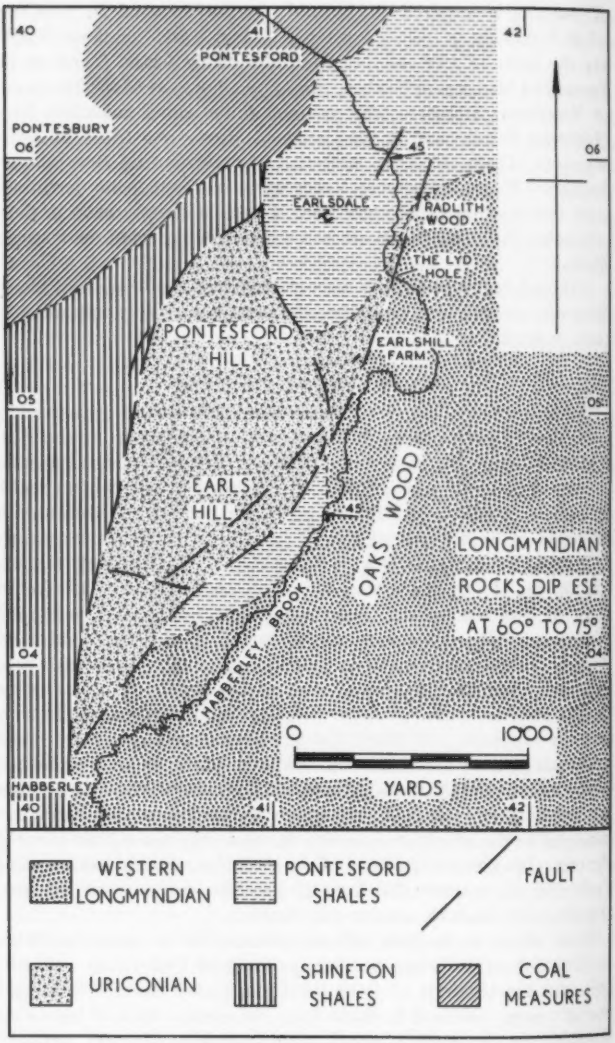
Although it has previously been claimed that the Pontesford Shales form one uninterrupted outcrop, the present mapping indicates that two are present.

(*b*) *The Radlith Wood district*

The base of the Pontesford Shales rests with marked unconformity upon the Uriconian rocks in the vicinity of the stream at and near Radlith Wood. The basal bed is a coarse, poorly-sorted, subangular conglomerate comprising pebbles of volcanic and sedimentary rocks, and quartz-pebbles. As far as exposures permit, the thickness is estimated to be between 10 and 13 feet. Next in the succession are 15 feet of dark-grey and black, pyritous and silty shales, poorly exposed and dipping almost due west at 45°. They are followed by 150 feet or more of buff-weathering, grey-green, thinly interbedded, silty and micaceous shales, mudstones and siltstones. At the top of this subdivision lies the exposure from which all the shelly fossils have been collected ; here the dip remains at about 45°, but the strike has swung slightly anti-clockwise. A fault with, apparently, only a small throw cuts the exposure, and above the last mentioned strata there occur grey shales, poorly exposed in the brook and so far yielding no fossils.

The base of the Ordovician rocks of the Radlith Wood district lies at a height of 400 feet O.D., or perhaps slightly lower ; this figure was measured in the stream course, and the base may rise a little towards the east. In contrast, the base of the Ordovician strata south-east of Earl's Hill lies at about 500 feet O.D., and the two outcrops are now thought to be discrete, separated by faulting.

There seems to be little reliable evidence for the direct southern continuation of the main northern outcrop of Ordovician strata to link with the outcrops of Oaks Wood, as shown on the Geological Survey's map. South of Earlshill Farm the western flank of the valley is marshy and has suffered considerable superficial slipping. Between Earlshill Farm and The Lyd Hole, however, the ground is high and flat, suggestive of an underlying " hard " formation rather than soft



TEXT-FIG. 1.—Geological map of the Pontesford District of Shropshire.

shales, and it seems significant that the stream has chosen to cut through hard Longmyndian sandstones rather than the supposed outcrop of soft Ordovician shales a few yards to the west. Because of this, together with the difference in height of the Ordovician basal unconformity, it is suggested that some structural discontinuity separates the outcrops of Pontesford Shales situated respectively north and south of The Lyd Hole.

(c) *The area south-east of Earl's Hill*

The well-known outcrop of Pontesford Shales south-east of Earl's Hill consists essentially of an exposure in the bed and sides of Habberley Brook, further exposures in the vicinity being both poor and rare. One interpretation of the geology has been put forward by Whitehead (1929 ; *in* Pocock *et al.*, 1938, pp. 92–4) but the position is still far from clear. The principal feature of interest claimed by Whitehead was the presence of a sheet of Ordovician rhyolite between the Pre-Cambrian rocks and the Pontesford Shales. Whittard (1952, pp.148–149) has commented on the difficulties attendant upon this interpretation. He has pointed out that the rhyolite, which is devitrified, has more characters in common with lavas of Uriconian age than with others of known Ordovician age, and has suggested that the rhyolite, instead of dipping west under the Pontesford Shales, might equally well dip to the east, beneath the Longmyndian rocks.

The writers' recent re-examination of the exposures gives no assurance that a continuous sheet of rhyolite does, in fact, exist beneath the outcrop of Pontesford Shales and dipping concordantly with them to the west at about 45°. What has been called the basal conglomerate of the Pontesford Shales is extremely irregular in thickness, apparently attaining up to 5 feet as far as present exposures indicate, compared with the 12 feet reported by Whitehead. There is a sharp transition from conglomerate to shale, an irregular bedding plane separating coarse, well-cemented conglomerate from silty shales.

The nature of the junction between the conglomerate and the sheet of rhyolite seems to be questionable, and we have not found a convincing exposure of indisputably solid rhyolite. On the other hand, many of the rhyolite boulders are large, measuring 2 to 3 feet in length and 1 foot or so square in cross-section. There are no exposures showing solid rhyolite in contact with Longmyndian rocks and in a small excavation made in the western bank of the stream, fine, greenish-grey conglomerate was seen to rest unconformably upon red Longmyndian sandstone.

On the basis of both natural exposures and the small excavation noted above, the impression gained is of an irregular and extremely poorly-sorted conglomerate, possibly of Caradoc age, containing both pebbles

and boulders of rhyolite and resting unconformably upon the Pre-Cambrian rocks. The existence of a discrete, solid sheet of rhyolite is not convincingly demonstrated ; isolated small exposures might reveal only part of a single boulder of rhyolite within the conglomerate. The individual boulders are clearly all of the same variety of igneous material, but this in itself is not unusual. It has also been argued that the rhyolitic material differs slightly from the local Uriconian rhyolites, but exposures of the latter are necessarily limited in south Shropshire, and the Habberley Brook material may have been derived from a different source, not now visible. The present re-interpretation no longer calls for a hypothetical phase of Ordovician vulcanicity which cannot be correlated with known Ordovician volcanic rocks in the Welsh Borders.

The southernmost limit of the outcrop of Pontesford Shales remains conjectural. Auger holes have failed to penetrate the boulder clay between Earl's Hill and Habberley, but on topographical evidence and from stream exposures it seems likely that the lenticular outcrop terminates at no great distance south-west of Oaks Wood.

(*d*) *The graptolitic fauna*

Numerous graptolites have been collected from two exposures, both comprising strata only a short distance above the base of the Pontesford Shales.

Locality A.—This was designated previously as Locality 2 (Whitehead, 1929, p. 121) and later as Locality 3 (Whitehead *in* Pocock *et al.*, 1938, p. 259). It is situated in the east bank of Habberley Brook about 150 yards north of The Lyd Hole. The following species were found :—

Climacograptus antiquus Hall
Diplograptus leptotheca Bulman
Diplograptus multidens (Elles and Wood)
Lasiograptus sp.

Locality B.—This is Locality 3 of Whitehead (1929, p. 121), numbered later by him as Locality 4 (Whitehead *in* Pocock *et al.*, 1938, p. 259). It comprises the bed and banks of Habberley Brook 400 yards south-east of the summit of Earl's Hill. The faunal list includes :—

Climacograptus antiquus Hall (large individuals)
Climacograptus brevis Elles and Wood
Orthograptus cf. *apiculatus* Elles and Wood
Orthograptus calcaratus var. cf. *vulgatus* Lapworth

Dr. Strachan has kindly identified all the specimens and he considers that both localities are in strata belonging to the Zone of *Diplograptus multidens*. The zonal graptolite has been found only at Locality A, but

Dr. Strachan points out that this form, the type-specimens of which were described from the Pontesford Shales, is of limited horizontal distribution and is not known with certainty in Shropshire outside the Pontesford district. Both localities are at approximately the same stratigraphical level, but the graptolites give no indication as to whether one may be slightly older than the other.

(*e*) *The shelly fauna*

Shelly fossils have been collected from only one locality in the Pontesford Shales, namely the eastern bank of Habberley Brook, 720 yards south-south-east of its junction with the Shrewsbury road, and 500 yards north of The Lyd Hole. This is Locality 1 of Whitehead (1929, p. 121 ; *in* Pocock *et al.*, 1939, p. 259). The most abundant and stratigraphically important fossils are the trilobites, but other groups are represented to a lesser degree. Unless otherwise stated, specimen numbers refer to collections in the British Museum (Natural History).

Family TRINUCLEIDAE Hawle and Corda
Subfamily CRYPTOLITHINAE Bancroft
Broeggerolithus broeggeri (Bancroft)
(Pl. XX, fig. 4)

1929. *Cryptolithus broeggeri* n. sp., Bancroft, p. 85, pl. 1, figs. 6–8.
1960. *Broeggerolithus broeggeri* (Bancroft), Dean, p. 111. This publication includes a full synonymy of the species.

B. broeggeri was first described from the Caradoc Series of the Onny Valley where it is abundant in, and restricted to, the upper Glenburrell Beds, belonging to the Lower Soudleyan Stage. The species is less common at Habberley Brook and only 10 cranidia have been collected. The state of preservation is not always good, but the pitting of the cephalic fringe, as well as the average size of individuals, appears to match closely that of the Onny Valley population.

Salterolithus caractaci (Murchison)
(Pl. XX, figs. 1, 5)

1839. *Trinucleus caractaci* Murchison, pp. 303, 650, 659, pl. 23, figs. 1 *a*–c only.
1957. *Salterolithus caractaci* (Murchison), Cave, p. 284, pl. 10, figs. 3–6.
1960. *Salterolithus caractaci* (Murchison), Dean, p. 96, pl. 15, figs. 1, 2, 4, 6, 9.

The type-locality of *S. caractaci* is Trilobite Dingle [= Bron-y-Buckley Wood], Welshpool, where the species is found in the Trilobite Dingle Shales of Harnagian age, but in the Onny Valley it occurs abundantly in the upper part of the Harnagian and then continues, though less commonly, into the Lower Soudleyan. In the upper Glenburrell Beds of Horderley the species occurs with *Broeggerolithus broeggeri* but in smaller numbers ; this is in contrast to the distribution

at Habberley Brook, where the great majority of the trinucleid population belongs to *S. caractaci*.

Family HOMALONOTIDAE Chapman

Brongniartella aff. *minor* (Salter)

(Pl. XX, figs. 3, 7)

This form of *Brongniartella* is known from only a few specimens, mainly fragmentary but including the external mould of a nearly complete dorsal exoskeleton, figured here as a latex cast. It is likely that a new specific or subspecific name may prove necessary when more material becomes available, but the Habberley Brook specimens are certainly identical with others collected from the Lower Soudleyan of the Onny Valley, where they have been recorded previously as *Brongniartella* sp. (Dean, 1958, p. 220).

Family ASAPHIDAE Burmeister

Subfamily ASAPHINAE Burmeister

Parabasilicus powisi (Murchison)

(Pl. XX, fig. 8)

1839. *Asaphus powisi* Murchison, p. 661, pl. 23, fig. 9*c*.

This oft-quoted Middle Ordovician trilobite species was first described, as *Asaphus powisi*, by Murchison (1839, p. 661, pl. 23, fig. 9) who employed two syntypes. One of these (*loc. cit.*, pl. 23, fig. 9 *a* and *b*) is now known to be a typical specimen of *Chasmops extensa* (Boeck) from the Onny Valley, and the other syntype (*loc. cit.*, pl. 23, fig. 9 *c* only), the external mould of a large pygidium with attached thorax, is chosen here as lectotype. The latter specimen, Geological Survey and Museum 6841, originated from the, presumably, Harnagian Stage of Trilobite Dingle, Welshpool, and so far no topotype cephalon has been illustrated. However, the lectotype can be matched with specimens from the Lower Soudleyan of the Onny Valley, and these, in turn, can be matched with the Habberley Brook specimen which has the cephalon attached. The species has been referred to the Korean genus *Parabasilicus* by Kobayashi (1934, p. 476). The distribution of *P. powisi* in the Welsh Borders is still imperfectly known, but specimens resembling the lectotype are known so far from the Harnagian and Soudleyan Stages.

Other shelly fossils

One specimen (Pl. XX, fig. 6), the internal mould of a dorsal valve, of *Reuschella* cf. *horderleyensis* Bancroft has been recovered from Habberley Brook. The species itself was described by Bancroft (1928, p. 187) from the Upper Soudleyan Stage, the Lower Horderley Sandstone, near Horderley. In the Caradoc district *R. horderleyensis* has proved to be of stratigraphical value for the Upper Soudleyan (Bancroft,

1929, table opp. p. 76 ; Dean, 1958, p. 204), and this new record is of particular interest as being the only one so far from older strata. A few small brachiopods of dalmanelloid aspect have also been collected from Habberley Brook. They are small and too poorly preserved for certain identification but bear a general resemblance to *Soudleyella avelinei* (Bancroft) from the Lower Soudleyan. The only other brachiopod found is a small inarticulate form, possibly referable to *Craniops* but indifferently preserved.

A few ostracods have been collected. All belong to the species described by Harper (1947, p. 347) as *Tetradella scripta*, but since referred by Henningsmoen (1953, p. 214) to *Tallinnella*. *T. scripta* is particularly abundant in the Harnagian Stage of the Caradoc district, east of The Longmyd, but has proved to be of little stratigraphical value as yet, and forms closely resembling the type have been found both as low as the Costonian Stage and as high as the Actonian Stage.

The only other shelly fossils collected are a few indeterminate gastropod fragments (possibly *Sinuites* sp.) together with occasional machaeridian plates. Single plates of *Lepidocoleus suecicus* Moberg are the more common, but one kite-shaped plate of *Plumulites* has also been found (Pl. XX, fig. 2). Similar machaeridian remains are fairly abundant in the Harnagian strata of the Caradoc district, but have proved of little or no stratigraphical use owing to their long vertical range.

(*f*) *Correlation with other successions*

There can be no doubt that the lower, graptolitic portion of the Pontesford Shales belongs to the Zone of *Diplograptus multidens*. The upper portion with its shelly fauna can be correlated accurately with the Lower Soudleyan Stage, Zone of *Broeggerolithus broeggeri*, in the Caradoc district and elsewhere in the Anglo-Welsh area, and it follows therefore that the graptolitic shales represent all or part of the Harnagian Stage. This provides additional confirmation of part of the correlation of shelly and graptolite zones proposed in east Shropshire (Dean, 1958, pp. 226–230). Unfortunately the total absence of shelly fossils from the lower strata makes it impossible to assess the proportion of the Harnagian present, but it is worth recording that Callaway's original correlation of the Pontesford Shales with the Harnage Shales was remarkably near the truth. The unconformity below the Pontesford Shales can be matched elsewhere in east Shropshire, though the break there occurs also below the *Nemagraptus gracilis* Zone, and the base of the Caradoc Series is diachronous. The well-known *Nemagraptus gracilis* Transgression was widespread throughout England and Wales but was also prolonged in time and continued through part of the succeeding *Diplograptus multidens* Zone.

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EXPLANATION OF PLATE XX

FIG. 1.—*Salterolithus caractaci* (Murchison). × 1·4. Partly exfoliated, nearly complete individual. In. 54296.

FIG. 2.—*Plumulites* sp. × 4·5. Internal mould of single kite-shaped plate. In 59750.

FIG. 3.—*Brongniartella* aff. *minor* (Salter). × 2. Internal mould of pygidium. In. 54302.

FIG. 4.—*Broeggerolithus broeggeri* (Bancroft). × 3. Incomplete cranidium showing characteristic form of cephalic fringe. In. 54292.

FIG. 5.—*Salterolithus caractaci* (Murchison). × 2. Internal mould showing underside of cephalic fringe. In. 54298.

FIG. 6.—*Reuschella* cf. *horderleyensis* Bancroft. × 2. Internal mould of dorsal valve. BB. 26164.

FIG. 7.—*Brongniartella* aff. *minor* (Salter). × 1·4. Latex cast from external mould of nearly complete dorsal exoskeleton. In. 54301.

FIG. 8.—*Parabasilicus powisi* (Murchison). × 0·7. Internal mould of thorax and cranidium with left librigena partly displaced. In. 54305.

All specimens are from the eastern bank of Habberley Brook, 720 yards south-south-east of its junction with the Shrewsbury road, near Pontesford, Shropshire. The horizon is in the upper portion of the Pontesford Shales, Soudleyan Stage, Zone of *Broeggerolithus broeggeri*. Specimen numbers refer to collections in the British Museum (Natural History).

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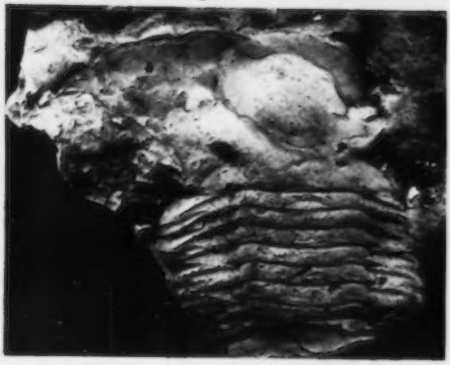

8

ORDOVICIAN FOSSILS FROM PONTESFORD.

facing p. 376.

The Antipodal Location of Continents and Oceans

By F. F. Evison and P. Whittle

Abstract

Five-sixths of all the continental area on the Earth's surface has antipodes in oceanic regions. This proportion is not significantly different from what one would expect on the assumption that the continents are randomly situated.

The fact that most of the land on the Earth's surface is antipodal to sea is often cited as a phenomenon of major importance, particularly in connection with the problem of the origin of the continents and oceans. From Lowthian Green's tetrahedral hypothesis of 1875 to the recent convection hypothesis of Vening Meinesz (1952), theories of the development of the Earth's surface have taken this antipodal relation as one of the salient features requiring explanation. Whether this aspect of the distribution of land and sea is significant from a statistical viewpoint does not seem to have been discussed. Umbgrove (1947) asserts that " the distribution of the continents is not haphazard but displays a certain regularity "; but since all land together occupies only about one-third of the Earth's surface it is not self-evident that the proportion of land situated antipodal to land is less than might be expected from a random distribution.

The antipodal relation is actually less pronounced than is commonly supposed. In the standard texts of Bucher (1941), Holmes (1944), and Umbgrove (1947), for example, it is stated that only one-twentieth of the land surface is antipodal to land. The correct figure is about one-eighth ; this can be quite easily checked, for virtually all the areas in question are to be found in South America, Antarctica, and their respective antipodal regions. This value refers to dry land, however, and requires some adjustment in the present context, since the continental shelves should be regarded as part of the continental area rather than the oceanic. On this basis Vening Meinesz (1952) has estimated that the " antipodal rule " applies to all but 6 per cent of the Earth's surface ; in other words, rather more than one-sixth of the continental area is antipodal to continent. Even this figure Vening Meinesz considers " too small to warrant a belief in a purely accidental distribution ".

Let us examine the matter by assuming an accidental distribution and calculating the expected amount of antipodal correspondence. Suppose that there are n continents, whose surface areas, expressed as proportions of the surface area of the Earth, are $s_1, s_2 \ldots s_n$. Suppose also that the positions of these continents are varied randomly and independently over the Earth. Then the probability that a given point is not covered by the jth continent is $1 - s_j$, and the probability that

neither the point nor its antipode is covered by the jth continent is $1 - 2s_j$. (We may assume that no continent is itself capable of covering any two antipodal points.) Hence the probability that a given point and also its antipode are covered by continent is given by

$$P_1 = 1 - 2(1 - s_1)(1 - s_2) \ldots (1 - s_n) + (1 - 2s_1)(1 - 2s_2) \ldots (1 - 2s_n).$$

P_1 is also the expected fraction of the Earth's surface having continent antipodal to continent.

The values of s for the existing continents may be calculated from the areas of continents and continental shelves given by Daly (1934) :

Eurasia . . .	0·131
Africa . . .	0·061
N. America . .	0·060
S. America . .	0·040
Antartica . .	0·029
Australia . .	0·023

Hence we find $P_1 = 0{\cdot}074$. On this basis we should expect to find 7·4 per cent of the Earth's surface with continent antipodal to continent. This estimate is not entirely realistic, however, for in order to make the problem amenable to theory we have tacitly assumed that continents may overlap one another.

Some idea of the effect of this assumption may be gained by calculating the theoretical probability that a given point is covered by continent. This is given by

$$P_2 = 1 - (1 - s_2)(1 - s_2) \ldots (1 - s_n)$$

which is also the expected fraction of the Earth's surface covered by continent. We find

$$P_2 = 0{\cdot}301$$

But the actual fraction of the Earth's surface covered by continent is

$$s_1 + s_2 + \ldots + s_6 = 0{\cdot}344$$

Our theoretical value P_2 is thus about 12 per cent too small. The theoretical value P_1 will also be too small, in about the same proportion. But the ratio P_1/P_2 is approximately correct. This represents the expected fraction of continental area that is antipodal to continent, and comes out at about one-quarter.

Summarizing, we find that whereas about five-sixths of the Earth's continental regions are antipodal to ocean, the statistical expectation for randomly distributed continents is about three-quarters. The difference between these values is so small that no special explanation of the antipodal location of continents and oceans seems to be required.

Whether the location of the continents is actually random is not susceptible of proof, but no evidence that a non-random influence has operated can be derived from the antipodal relation.

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The Central and Eastern Parts of the Lammermuir-Stranraer Moraine

By J. B. Sissons

Abstract

The Lammermuir-Stranraer moraine was first described by Charlesworth and interpreted by him as marking an important readvance. The central and eastern parts of this feature are discussed. It is concluded that it is not a true moraine in these areas and that the readvance supposed to have formed it did not occur.

I. Introduction

In 1926 J. K. Charlesworth described a kame-moraine extending along much of the northern edge of the Southern Uplands from St. Abb's Head in the east to Stranraer in the west. He named this feature the Lammermuir-Stranraer moraine and interpreted it as marking the limit of a major readvance of Highland ice. In another paper, published at the same time, he correlated a readvance identified in the western part of the Southern Uplands (the Kirkcowan stage) with the kame-moraine. The term Lammermuir-Stranraer moraine has also come to be applied to the combined limits of Southern Upland and Highland ice at this stage.

The moraine has been subsequently accepted by most writers concerned with glacial retreat stages in Scotland (and in Britain as a whole) as an important stage and has been variously correlated with stages in continental Europe. It appears, for example, in maps by Wills (1951) and Synge (1952) of the major retreat stages in Britain. Manley (1951) discussed the climate at the time the Lammermuir-Stranraer moraine was formed. Movius (1942) referred to it in his " Irish Stone Age " while Lacaille (1948) accepted it in his paper on the deglaciation of Scotland in relation to early man's environment. It appears in maps and text in works on pollen analysis such as Godwin's " History of the British Flora " (1956) and Donner's paper on Late-glacial retreat stages in Scotland (1957). Among important foreign books in which it appears are Flint's " Glacial and Pleistocene Geology " (1957) and Woldstedt's " Das Eiszeitalter " (1958). Recently Charlesworth has repeated his interpretation in summary form in " The Quaternary Era " (1957, pp. 1204–5).

In view of such authoritative and widespread acceptance of the Lammermuir-Stranraer stage one almost hesitates to point out that the conclusions of almost all local workers other than Charlesworth are opposed to the existence of such a stage. Some of the evidence on which these conclusions are based was published before the appearance of Charlesworth's paper on the Lammermuir-Stranraer readvance and in that paper he discussed this evidence and interpreted it differently.

Nevertheless, subsequent work has often demonstrated the validity of the conclusions of the early workers.

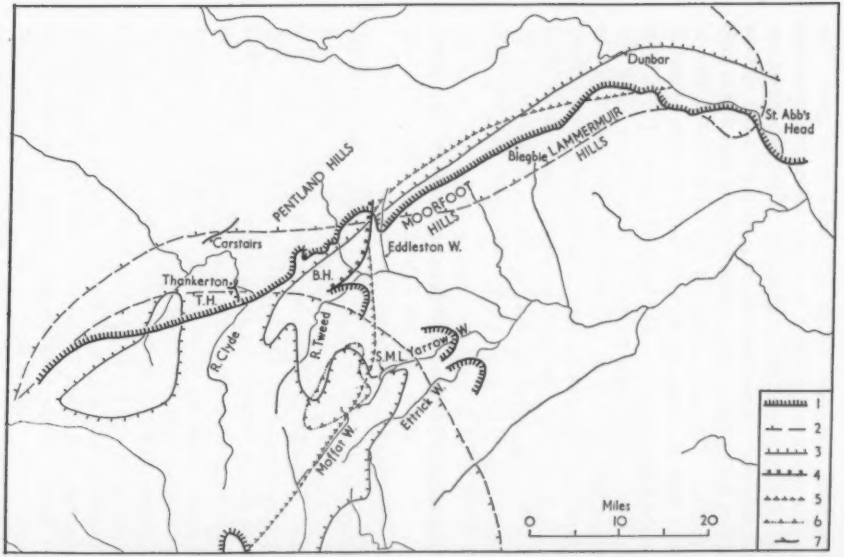
Since Charlesworth (1926b) described the moraine as best developed in its central and eastern parts, and since the present writer has studied these in some detail, this paper will be restricted to these parts. The area concerned is shown in Text-fig. 1, in which various interpretations of the Lammermuir-Stranraer limit are marked. These interpretations have been taken from maps published by the authors indicated with the exception of that due to Bailey and Eckford, which is described in their text (1956, p. 260). The most significant line is, of course, that of Charlesworth (1926b, map opp. p. 50) and this is the main subject of discussion in the following pages. The limits of Movius (1942), Wills (1951), and Donner (1957) were published on small-scale maps and therefore minor differences from Charlesworth's line should not be considered significant being, presumably, attributable to cartographic limitations. It is not clear, however, on what evidence these limits were based where they differ significantly from Charlesworth's, for all three authors refer to his work but do not explain why they have modified his interpretation.

II. The Lammermuir-Stranraer Moraine Between St. Abb's Head and the Clyde

(a) *From St. Abb's Head to Blegbie*

The Lammermuir Hills rise to an altitude of about 1,700 feet in the western part of this area and decline eastwards to about 700 feet near St. Abb's Head. The hills are limited northwards by a steep slope, the base of which is often about 750 feet above sea-level in the west. Through much of this distance the lower slopes of the Lammermuir scarp and the more-gently-inclined ground at its foot are mantled with fluvioglacial sand and gravel, which locally extends up to 1,000 feet or so, but is mainly concentrated below about 750 feet. In some places this material forms ill-defined spreads with no definite form, but very frequently it occurs as well-marked kames, kame terraces and eskers, sometimes associated with kettles. These features and the numerous meltwater channels that occur in the same area were described by Kendall and Bailey (1908). Charlesworth grouped the depositional forms together as the eastern part of the Lammermuir-Stranraer moraine. This interpretation meets several difficulties, including the following.

(i) The moraine is almost entirely composed of fluvioglacial material. Where till does occur in the moraine it appears to be restricted to small patches on ice-contact slopes and small areas of ablation moraine of limited thickness. (The morainic mounds with blocks of Old Red Sandstone referred to by Kendall and Bailey (1908, p. 19) are actually



TEXT-FIG. 1.—The eastern and central parts of the Lammermuir-Stranraer moraine according to 1, Charlesworth (1926b, map opp. p. 50); 2, Movius (1942, map p. 52); 3, Wills (1951, map p. 62); 4, Bailey and Eckford (1956, p. 260); 5, Donner (1957, map p. 223). 6, Generalized limits of the valley glaciation in the area between the Yarrow and upper Tweed, partly after Young (1864). 7, The Carstairs and Thankerton "kames". T.H. = Tinto Hills. B.H. = Broughton Heights. S.M.L. = St. Mary's Loch.

an esker, the gravels of which are in places covered by up to 2 feet of angular ablation moraine.) No constructional mounds composed largely or entirely of till have yet been found. This evidence does not support the interpretation of the features as an end-moraine formed by a major readvance.

(ii) The view that the belt of fluvioglacial features is a moraine implies that these features are of approximately the same age throughout the length of the belt. Detailed mapping and interpretation of these features (and of the meltwater channels) shows, however, that they were not formed contemporaneously, but are older in the east than the west (Sissons, 1958b, p. 73). The ice-margin lay at a slight angle with the fluvioglacial belt during the deposition of the latter, as is shown, for example, by well-marked kame terraces alternating with channels in aligned sequence.

(iii) Although Charlesworth stated that the Lammermuir-Stranraer moraine " marks the limit of a readvance of uncertain yet probably considerable magnitude " (1926b, p. 44), he also stated that this moraine lies to the north of the line of maximum readvance in the Lammermuir area (1926b, p. 28). This seems a rather serious anomaly in his interpretation. His conclusions were based on Kendall and Bailey's (1908) valid evidence that the meltwater channels imply that the last ice to occupy East Lothian reached an altitude of over 1,000 feet on the western part of the Lammermuir scarp, but that the main concentration of fluvioglacial deposits is around the scarp foot, especially below about 750 feet. This concentration owes its existence and location to the concentration of meltwater flow in the depression between the scarp face and the ice mass wasting back from it. Into this depression flowed streams from the Lammermuir scarp and from the wasting ice. At times the marginal zone of the ice was buried beneath the fluvioglacial deposits resulting in the kame-and-kettle topography that is locally well developed. The influence of the pre-existing relief on deposition is shown, not only by the abundance of the deposits along the scarp foot for much of its length, but also by the local concentration of material in pre-existing valleys, at the base of steep slopes and at points where streams descending from the Lammermuirs were checked in their rapid progress by the ice mass and caused to deposit their heavy loads. The abundance of fluvioglacial deposits in this area in fact seems to suggest, not a readvance of the ice, but a period of rapid glacier decay.

It is therefore concluded that the fluvioglacial belt following the foot of the Lammermuir scarp is not a moraine and is not related to a halt or readvance of the ice-margin.

Whether or not the higher line marking the upper limit of definite glacial retreat features on the Lammermuir scarp is significant has yet to be ascertained. This line falls gently eastwards from an altitude of

about 1,200 feet in the western part of the area now being considered. It is tempting to conclude that the last major readvance of Highland ice failed to surmount the Lammermuirs and that the upper limit of definite retreat features marks its maximum extent. Hollingworth (1952), however, has pointed out that conclusions of this type may not be valid, stating that " It is generally accepted that marginal drainage will only operate below the snow line ". On the Lammermuir scarp additional factors preventing the formation of channels and related deposits on the upper slopes might have been the steepness of these slopes and the small available drainage areas of streams from the ice-free area. The problem might be solved by studying solifluction deposits, for it is a reasonable expectation, especially in an upland area such as the Lammermuirs, that such deposits would be strongly developed outside the limit of a readvance, but not at the surface within it (having been destroyed by the readvance or buried by its deposits). Such studies have not been carried out in the Lammermuirs and hence the significance of the upper limit of definite retreat features remains uncertain.

(*b*) *From Blegbie to the Eddleston Valley*

For some 17 miles from Blegbie farm to the Eddleston valley the Lammermuir-Moorfoot plateau (*c.* 1,200–2,000 feet) ends in a nearly-straight scarp facing north-west. The ground near the scarp foot is usually an area of subdued relief at between 800 and 1,000 feet altitude. Abundant fluvioglacial deposits occur in this latter area, in places forming complex areas of kames, eskers, and kettles. These deposits and forms continue southwards into the Eddleston valley. All these deposits were considered by Charlesworth to form part of the Lammermuir-Stranraer moraine produced when the margin of the Highland ice lay along the foot of the Lammermuir-Moorfoot scarp with a tongue extending southwards down the Eddleston valley (Text-fig. 1).

Among the difficulties encountered by this interpretation are some already mentioned in discussing area (a), namely, the absence of till moraines and the abundance of fluvioglacial material, evidence favouring rapid deglaciation rather than a period of readvance. A more serious difficulty is met with in this area, however, and will be considered first in relation to the Eddleston valley.

In this valley there is clear evidence that the last glacier ice came, not from the highlands, but from the Southern Uplands. For example, some 200 meltwater channels mapped by the writer in the area drained by the Eddleston Water (1958a, map p. 162) show that meltwaters flowed in a general south–north direction on both sides of the valley, turning north-eastwards to the north of the Moorfoot scarp. The inclination of the channels as a group reflects the direction of slope of

the ice surface and indicates that in the Eddleston valley area the ice came from the Southern Uplands.

Although striae are rare in this area, the only two marked on Geological Survey One-Inch sheet 24 indicate ice movement from south-west to north-east. Further evidence is provided by several eskers, up to a mile long, that occur immediately north-east of the northern part of the Eddleston valley. It is generally accepted that eskers (excluding Mannerfelt's (1945) subglacially-engorged type) often indicate the latest direction of former ice movement. The eskers near the Eddleston valley were formed by streams flowing south-west to north-east and thus imply that the last ice movement was in this direction.

Another line of evidence is provided by indicator stones. In fluvioglacial gravels in the Eddleston valley felsite from the Tinto Hills, which lie 20 miles to the south-west, has been found (Bailey and Eckford, 1956, p. 255). Furthermore, these gravels are free from coal fragments. The valley is cut in Ordovician and Silurian rocks, but Carboniferous rocks outcrop extensively immediately to the north and north-west of its head. Had the gravels been laid down by streams draining Highland ice they could hardly have failed to contain coal fragments. That the deposits were laid down by streams draining Southern Upland ice is also demonstrated by their bedding, which shows that these streams flowed northwards along the valley side or into the ice towards the valley axis.

There is thus ample evidence that the last ice to occupy the Eddleston valley came from the Southern Uplands. How far to the north this ice extended is not definitely established, but McCall and Goodlet's (1952) study of indicator stones strongly suggests that it reached at least as far as a point 8 miles due north of the head of the valley. One would, therefore expect that it would have also extended along the Moorfoot-Lammermuir scarp-foot zone for a considerable distance north-east of the Eddleston valley. The meltwater evidence favours this view.

In a zone up to several miles wide to the north-west of the scarp a considerable proportion of the meltwater features appear to have been formed subglacially. Proof of subglacial drainage is provided by a number of eskers, the most common direction of the longer ones being south-west to north-east. Several of these eskers can be traced into meltwater channels, strongly suggesting that these channels are of subglacial origin also. There are also several channels with up-and-down long profiles probably formed by meltwaters flowing subglacially under hydrostatic pressure. Throughout this area there is a marked contrast with the corresponding belt at the foot of the Lammermuir scarp north-east of Blegbie. In the latter area kame terraces and channels running along north- and north-west-facing slopes are numerous and

point to an important marginal and submarginal drainage developed in relation to the wasting Highland ice. In the Eddleston-Blegbie belt, however, kame terraces and channels running along slopes are relatively infrequent and most slopes facing north and north-west have no such features at all. On the other hand, several channels occur on slopes facing south and south-east. Furthermore, the south-eastern edge of the main gravel concentration is often some distance from the scarp foot, unlike the gravel belt in area (a), thus again suggesting a difference in origin.

On the scarp slope itself there is an even greater contrast between the parts north-east and south-west of Blegbie, In the former area a great number of gently-inclined channels and kame terraces run along the middle and lower parts of the slope. South-west of Blegbie, however, over a distance of 14 miles there are neither kame terraces nor channels along the scarp slope.

These marked contrasts can be explained by the former presence of Southern Upland ice in the area south-west of Blegbie. The absence of marginal and submarginal features on the scarp slope and on many other slopes facing north-west and north is then accounted for by the ice surface having been inclined away from the scarp (probably at a slight angle). The drainage of this ice mass was partly subglacial, as shown by evidence referred to above. The clear preservation of the eskers with their numerous minor bends shows that, at least in the later stages of deglaciation, the ice was stagnant. Stagnation is also evidenced by the numerous irregular masses of sand and gravel, often bounded by ice-contact slopes. It was these and other fluvioglacial deposits (including the eskers) that Charlesworth grouped together as a portion of the Lammermuir-Stranraer moraine formed along the margin of the Highland ice.

(*c*) *The area between the Pentland Hills and the Broughton Heights*

In the relatively low ground between these two upland areas are numerous fluvioglacial deposits with typical ice-contact forms included in the Lammermuir-Stranraer moraine by Charlesworth. Eckford (1952) has shown that these deposits were laid down in association with Southern Upland ice. As in area (b), the form and distribution of the deposits are indicative of dead ice.

(*d*) *The middle Clyde area*

In this area are two important linear accumulations of fluvioglacial material. The more northerly and larger is well known as the " Carstairs kame " and the other is referred to in the literature as the " Thankerton kame ". Charlesworth (1926b) interpreted each of these " kames " as an end-moraine formed along the margin of the Highland ice.

The Thankerton "kame" has been recently discussed in detail (Sissons, 1961) and will be considered only briefly therefore. It consists of several steep-sided fluvioglacial ridges situated in the valley of the Clyde east of the Tinto Hills (Text-fig. 1). Although differing as regards some aspects of its origin, Charlesworth (1926b) and other writers have all agreed that these ridges were formed partly or wholly along an ice-margin. There is clear evidence, however, that all the ridges are eskers. The principal evidence is as follows. (i) The ridges have steep ice-contact slopes on both sides and are in places margined by kettles, showing that ice was present on both sides of them when they were formed. (ii) The ridges have typical esker form with winding courses, and undulating crest lines, the highest points often occurring at bends. (iii) Two of the principal ridges can be traced back into valleys leading down from the Tinto Hills and the third main ridge can be followed back into a system of meltwater channels and other fluvioglacial features running along the northern side of these hills. (iv) The high proportion of felsite in the ridges shows that they were formed by streams coming from the felsite mass of the Tinto Hills, not from Highland ice that had come up the Clyde valley. The Thankerton ridges therefore do not indicate the former position of the margin of any ice mass. That the last ice to occupy the region in which they occur came from the Southern Uplands is indicated by striae (Geikie, 1873, p. 42), the distribution of Tinto felsite (McCall and Goodlet, 1952) and the numerous meltwater channels and fluvioglacial deposits of the surrounding region.

The Carstairs "kame" (Text-fig. 1) comprises a system of steep-sided fluvioglacial ridges, in places over 50 feet high, with a general south-west to north-east trend. The main ridge system extends over a distance of about 4½ miles and varies in width from a few tens of yards to about half a mile. The ridges split and rejoin one another but generally trend approximately parallel with the system as a whole. Between the ridges are peat-filled, often-closed depressions. South of the ridge system is an area of broad mounds, often with smooth slopes of gentle to moderate inclination. This latter area is dominantly sandy whereas the ridges comprise material ranging from sand to boulders over 2 feet across.

The composition and bedding of the mounds indicate deposition in a lake or lakes, as concluded by Gregory (1915, p. 473), Charlesworth (1926b) and Macgregor (1927, p. 498). The moundy form of these deposits and the irregular, sometimes-closed depressions between indicate deposition in association with dead ice.

The ridge system has been the subject of some dispute in the literature. Charlesworth (1926b) concluded that it was deposited by streams draining Highland ice. The majority opinion, however, favours an

origin in relation to Southern Upland ice (e.g. Geikie, 1894, p. 187; Gregory, 1915; Macgregor, 1927). There seems to be no reason for adopting Charlesworth's minority interpretation: he himself referred (1926b, p. 35) to the " apparently overwhelming evidence " against it. Particularly significant is the fact that almost all the material either has been derived or could have been derived from a southerly or south-westerly source. For example, Gregory (1915, p. 471) found in one section 30 per cent of greywackes and other Southern Upland sedimentary rocks but only 2 per cent of Highland rocks. In order to overcome this difficulty Charlesworth suggested various possibilities, the principal one being that the presence of Southern Upland erratics is the result of Highland ice having advanced over ground previously covered by Southern Upland ice. It seems much more likely, however, that the high proportion of Southern Upland erratics is due to Southern Upland ice having advanced over ground earlier occupied by Highland ice. Further east Eckford (1952) and McCall and Goodlet (1952) have shown that this in fact occurred (sections reveal Southern Upland deposits overlying Highland ones) and a similar sequence of events is therefore expectable in the Carstairs area.

Charlesworth and Gregory both concluded that the Carstairs ridge system was formed along an ice margin by streams draining from the ice. This conclusion, however, cannot be reconciled with the form of the feature. Streams draining from an ice mass would have produced either alluvial fans sloping away from the ice or flat-topped deltas. Yet the Carstairs ridges have narrow, undulating crest-lines and are steep on both sides. These steep slopes and the numerous associated kettles show that ice lay on both sides of each ridge when it was formed. This was recognized by Macgregor, for he suggested (1927, p. 498) that the ridges " may have been laid down between ice-walls ". Macgregor also suggested, however, that the ridges were laid down between dead Southern Upland ice to the north and active Southern Upland ice to the south. This interpretation seems rather improbable: in fact the numerous kettles and the frequent bends in the well-defined ice-contact slopes of the ridges indicate that the ice on both sides of the ridge system was stagnant. Macgregor's interpretation also implies that the system was formed in an open ice-walled channel. This is probably partly true, but at least in the north-eastern part a subglacial origin seems more probable. Here there is one main ridge for two miles with a smaller ridge (joined to it at both ends) over a third of this distance. These two ridges, with their steep sides, undulating crests and winding courses, appear to be typical subglacially-formed eskers. Referring to the Carstairs ridges as a whole, Gregory (1915, p. 473) remarked how they " strikingly resemble some typical Swedish eskers " and one can also compare them with some of the Canadian eskers with their

numerous sub-parallel ridges and associated steep-sided kettles. The Carstairs ridges are, in fact, a complex esker and, like the Thankerton esker system further south, form part of the extensive glacial drainage system of the Southern Upland ice sheet, which, over a large area extending from the borders of East Lothian southwards to the source of the Tweed and westwards beyond the Tinto Hills, trends towards a direction between north-east and north.

Thus the concept of a Lammermuir-Stranraer moraine formed along the margin of Highland ice finds no support in the middle Clyde area. The two sets of features in this area interpreted by Charlesworth as end-moraines formed by this ice are both eskers (or esker systems) formed in association with Southern Upland ice.

III. The Lammermuir-Stranraer Stage in the Tweed Basin

As mentioned above, Charlesworth (1926a, 1926b) described and mapped the limits of the Southern Upland ice at the Lammermuir-Stranraer stage. In the vicinity of the Solway he showed this limit as a continuous line, but in the Tweed basin he indicated only the termini of three valley glaciers (1926b, map opp. p. 50). In his text he stated that these glaciers occupied the valleys of the Yarrow and Ettrick Waters and of the upper Tweed. He said that at the same time corrie glaciers existed in the Lammermuirs. In none of these areas, however, are there moraines or other evidence to support this interpretation.

Movius (1942) interpreted the Scottish Readvance described by Trotter (1929, pp. 600–602) in the Carlisle area, Charlesworth's Lammermuir-Stranraer Readvance, and the Perth Readvance of Simpson (1933) as representing a single stage. Other writers, however have considered these three to be distinct and successive stages. Movius's reasons for differing so radically from other interpretations are not given. Furthermore, he stated (p. 53): " Actually the limit of this ice from the Southern Uplands has not been traced north of Brampton in Cumberland . . .". The line shown on his map does not correspond with any evidence in the Southern Uplands indicating the limits of a readvance.

A novel feature of a map included by Wills in his Palaeogeographic Atlas (1951, p. 62) is the delimitation of a nunatak area of considerable extent at the Lammermuir-Stranraer stage in the Southern Uplands. The present writer is not aware of any evidence that suggests the former existence of such a nunatak. Wills' Lammermuir-Stranraer limits in the Southern Uplands correspond approximately with moraines only in part of the area between the Yarrow Water and upper Tweed. These moraines are referred to below.

In a fairly recent paper Bailey and Eckford (1956) suggested a modi-

fication of Charlesworth's interpretation in the Eddleston valley and adjacent parts of the Tweed valley. They concluded that the edge of the Southern Upland ice lay along the western slope of the Eddleston valley and continued south-westwards into the Tweed valley (Text-fig. 1). This interpretation, which has been recently republished in summary form (1960) is opposed to that of the present writer (1958a). The main evidence used by Bailey and Eckford is the fluvioglacial deposits. In the Eddleston valley these deposits are concentrated on the western side (in the area between their ice-margin shown in Text-fig. 1 and the Eddleston Water). Bailey and Eckford interpreted the deposits as formed by proglacial streams that flowed from the ice down the western side of the Eddleston valley. The forms of these deposits show conclusively, however, that they were laid down in intimate association with dead ice, for they occur as kames, eskers, kame terraces and kettles. These forms and their bedding show that the deposits were washed into the ice, not out of it (Sissons, 1958a). This is also demonstrated by about 100 meltwater channels that occur on the valley side immediately above the deposits. These channels, none of which was mentioned by Bailey and Eckford, mostly slope down northwards or towards the axis of the valley and could have been cut only when ice occupied the valley. The relations of the channels to the deposits show that the streams that cut the former also laid down the latter. In the Tweed valley also the evidence of meltwater channels and fluvioglacial deposits is not in accord with Bailey and Eckford's interpretation (R. J. Price, unpublished).

The most recent interpretation of the position of the Southern Upland ice-margin at the Lammermuir-Stranraer stage is shown in a map by Donner (1957, p. 223). In the Eddleston valley his line corresponds in part with Bailey and Eckford's line already discussed (Text-fig. 1). Further south it crosses the Tweed at an altitude of about 600 feet, rises to about 2,000 feet on the Tweed/Yarrow watershed and then descends to 800 feet near St. Mary's Loch. Such an ice-margin position seems very improbable and study of the area reveals no evidence to justify it. Southwards, however, Donner's line does correspond in part with a series of readvance moraines. Some of these moraines, which are composed mainly of till, often with large angular blocks, were described by Young in 1864. The area in which moraines occur is shown in generalized form in Text-fig. 1. As Young clearly demonstrated, these moraines were produced by local valley glaciers (it is not possible to show these individually in Text-fig. 1) and this group of moraines is restricted to the valleys of the high hill area between the Yarrow Water and upper Tweed. Similar moraine areas, not shown in Text-fig. 1, occur further west in the Southern Uplands in and around areas of high ground. These moraines, however, represent a valley glaciation

(or glaciations) of limited extent and cannot be integrated to form part of an extensive ice-sheet as implied by Donner's map.

IV. Conclusion

It has been shown that the central and eastern parts of the Lammermuir-Stranraer moraine are composed almost everywhere of fluvioglacial deposits formed in association with dead ice. Over a considerable length of the northern edge of the Southern Uplands the last glacier ice came from the Southern Uplands, not from the Highlands as Charlesworth maintained. The only true moraines occur in the higher parts of the Southern Uplands, but these represent a local valley glaciation or glaciations. Only in the Lammermuir area did the last ice come from the Highlands and it is not yet known if the limits of a readvance of this ice were reached in this area.

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Erosion of a Rock Floor by Tidal Sand Streams

By D. T. Donovan and A. H. Stride

(PLATE XXI)

Abstract

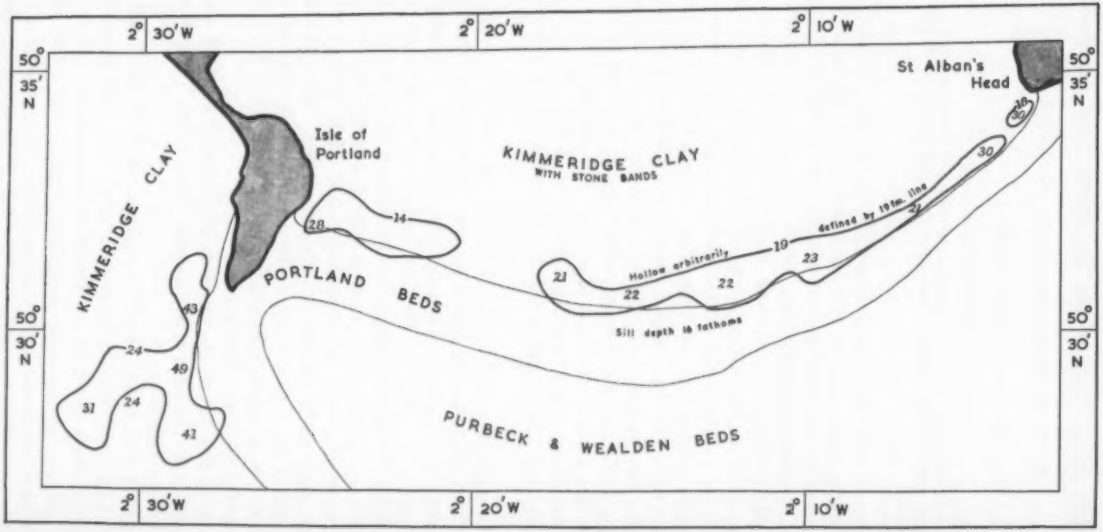
Some closed basins extending along the strike of soft rocks in the English and Bristol Channels were cut by sand moved by tidal currents.

The importance of tidal scour between islands has been reviewed by Kuenen (1950). More recently, attention has been drawn to the streams of sand inferred to be moving along the open continental shelf in the vicinity of southern Britain (Stride, 1959). The amounts of material being moved are large. For example, it has been estimated that in the southern part of the North Sea four million cubic metres must cross a plane 40 miles wide each year (Stride and Cartwright, 1958). It would be surprising if there had not been some erosion of the rock floor over which such sand has passed during the last 6,000 years, while the sea level has been approximately the same as to-day. Such is suggested as the origin of some closed basins, which extend along the strike of soft rocks and are bounded by harder ones. The geology of the two regions considered was mapped on cruises of R.R.S. *Discovery II* during 1959 and 1960, detailed accounts of which will be published elsewhere.

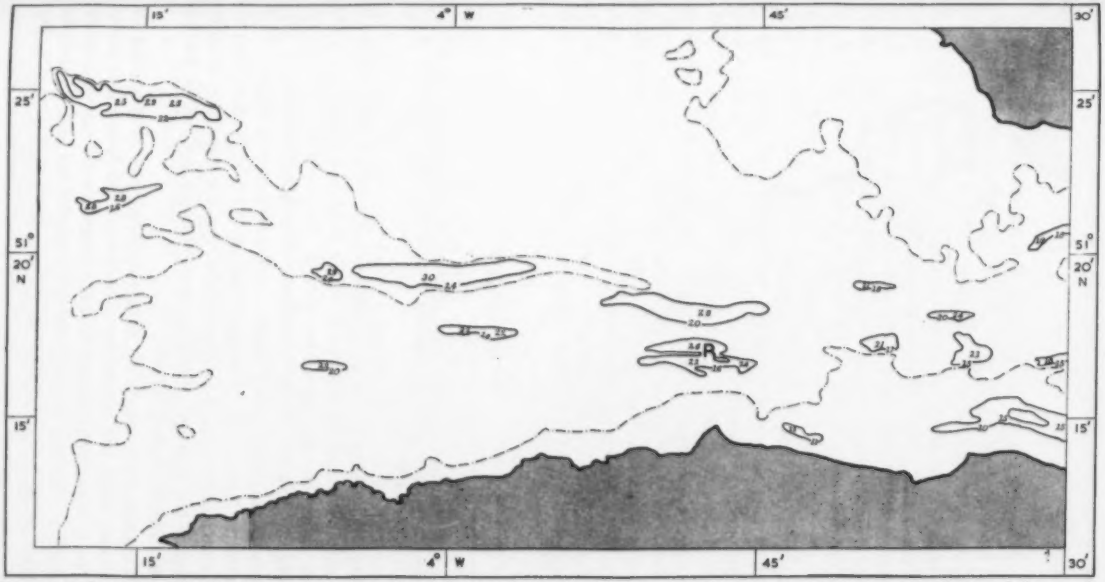
The first region lies off the coast of Dorset (Donovan and Stride, in press). The basins follow the line of outcrop of the upper Kimmeridge Clay from St. Alban's Head to the east side of Portland, a distance of nearly 15 miles (Text-fig. 1). They are flanked on the north and west by underlying Kimmeridge rocks, rich in resistant stone bands. The other side is defined by the up-standing ridge of even tougher Portland rocks. The deepest basin lies about half a mile off St. Alban's Head where soundings of 30 fathoms have been recorded. The 19-fathom contour defines this " deep " most clearly. Further from the coast the depth below the surrounding floor lessens until it is only about 5 fathoms lower than the sill of Portland Stone.

A similar hollow lies off the west side of the Isle of Portland where the greatest depth recorded is 49 fathoms. This " deep " is enclosed by the 24-fathom contour.

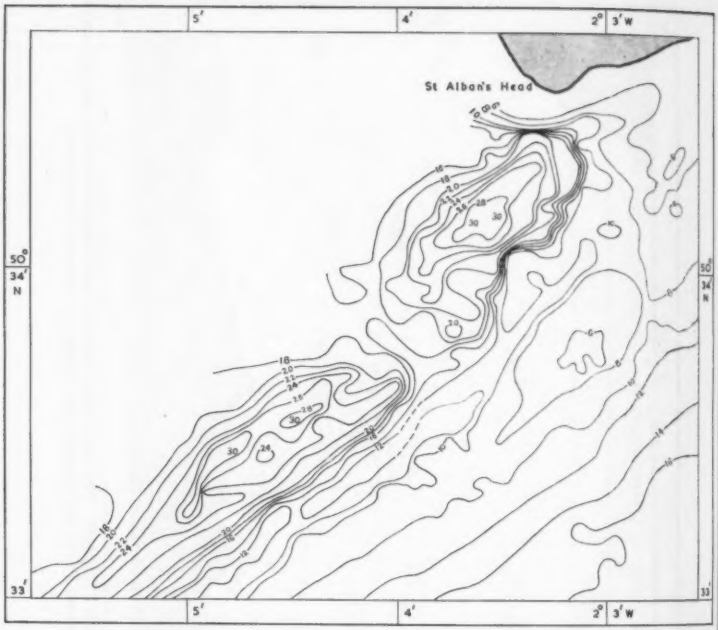
Hollows aligned along the strike are also numerous in the Mesozoic rocks of the Bristol Channel (Donovan, *et al.*, 1961; full account in preparation). Here also they are present in different sizes. There are shallow hollows between close-spaced stone bands and deeper ones where resistant bands are more widely spaced. The positions of the principal hollows there are shown in Text-fig. 2, the maximum depth



TEXT-FIG. 1.—Simplified geological map for sea floor south of Dorset showing strike-determined hollows cut in uppermost Kimmeridge Clay. The shallowest contour enclosing each hollow (or one defining it) and the maximum depth are shown in fathoms.



TEXT-FIG. 2.—Closed basins in the outer part of the Bristol Channel. The shallowest contour enclosing each hollow and the maximum depth are given in fathoms. The 10- and 20-fathom contours are shown by the usual convention.



TEXT-FIG. 3.—The eastern end of the strike-determined hollow below St. Alban's Head, Dorset. Contours above 6 fathoms not shown. Based on part of Admiralty Survey No. E8891.

and the shallowest contour which encloses the hollow being shown in each case. Some of the hollows lie in the floor of a submerged valley system, others are not associated with it. East of the area shown in Text-fig. 2 there is a hole 20 fathoms deep lying immediately west of Steep Holm island and 2 cables north-north-west of Rudder Rock (Admiralty Chart 1152). Admiralty Survey No. E5678 shows that the hole is enclosed by the 11-fathom contour and the sides have gradients as steep as one in three. We have not examined this hollow but it is almost certainly eroded in rock.

Both areas here described were doubtless dry land during the Pleistocene when the sea fell to several hundred feet below present level, so it might be supposed that the hollows were due to subaerial denudation. This can be ruled out in Dorset for the hollows there are unrelated to the land drainage and quite unlike subaerial topography, and the area was well beyond the furthest extension of the Pleistocene

ice sheet. They end abruptly against present day cliffs (Text-fig. 3, for example) and so must result from submarine erosion, with sand as the abrasive.

These arguments are less conclusive in the Bristol Channel. The hollows which lie in the submerged valley could have resulted from differential river-bed erosion, while others could have been gouged-out by the Irish Sea Ice which reached the northern shores of the Channel from seaward (Pringle and George, 1948, Fig. 27, p. 91). Nevertheless such origins cannot account for the very close relationship between relief and minor geological structure which is revealed by the asdic records, as illustrated in Plate XXI. The floor must have been modified, at least, by marine erosion.

It is probable that in both localities waves as well as tidal streams have been responsible for erosion, especially during periods when the sea was shallower than to-day. The overwhelming importance of the tidal streams off the Dorset coast is shown by the geographical correspondence between the deepest hollows and streams which, as measured at the sea surface, are several knots faster there than over the adjoining ground (Sager, 1961). Associated turbulence is made obvious by the patches of cloudy water reaching the surface, even when there are no sea waves. In the surface waters of the Bristol Channel tidal currents reach 3 to 4 knots in the area under discussion and sand transport must occur.

Conclusions

Tidal streams transporting sand can erode a rock floor. Closed basins south of Dorset and in the Bristol Channel are a result of this process. Similar erosion features may be expected elsewhere, formed under existing conditions or during lower or higher sea-level. Features now on dry land may have originated in this way during high sea levels of the Pleistocene.

The authors are indebted to Dr. J. N. Carruthers for helpful discussion.

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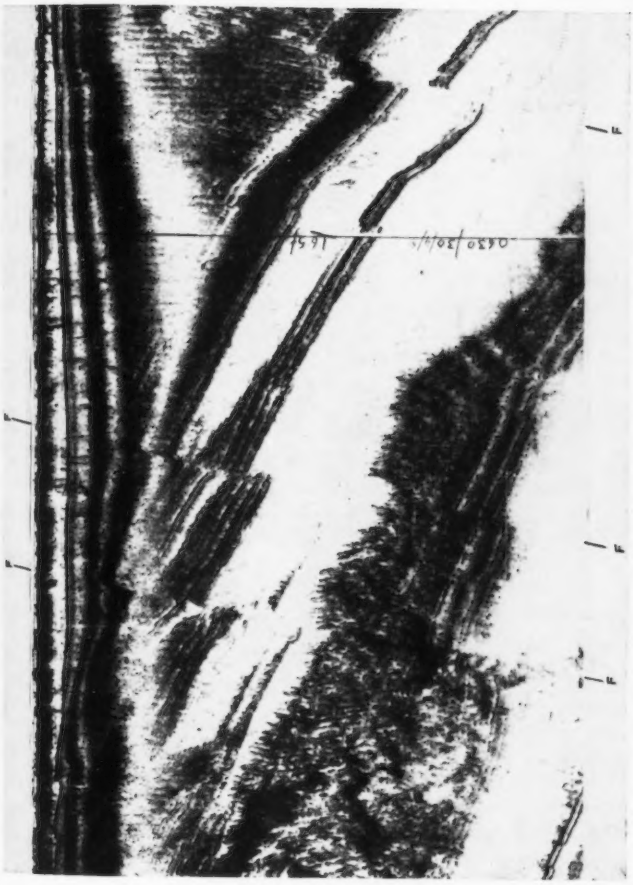
EXPLANATION OF PLATE XXI

Acoustic map (2½ miles × ½ mile) from the neighbourhood of "R" on Text-fig. 2. A faulted ridge (bottom left to top right) separates two hollows extending parallel to the strike. The thin dark lines are the trace of individual beds of hard rock separated by hollows left by the removal of clays.



ACOUSTIC MAP OF PART OF BRISTOL CHANNEL FLOOR

facing p. 398.

A Note Concerning the Age of the Metamorphism of the Dalradian Rocks of Western Ireland

By J. F. Dewey

Abstract

Evidence is adduced to relate the published minimum absolute date of − 475 m.y., for the metamorphism of the Connemara Schists, to its correct position in the Lower Palaeozoic stratigraphy of Western Ireland. The metamorphism pre-dates the *Didymograptus extensus* zone of the Arenig in Connemara and South Mayo and an Upper Cambrian or Tremadocian age is considered likely.

I. Introduction

The age of the Connemara Schists has been a source of controversy since the publication of Murchison's *Siluria* in 1859 and the age of the metamorphism affecting these rocks an even greater point of argument. During the second half of the nineteenth century there were two schools of thought concerning these problems. One, first outlined by Kinahan (1874), attributes great importance to the Ordovician/Silurian unconformity south of Lough Nafooey (Text-fig. 1).

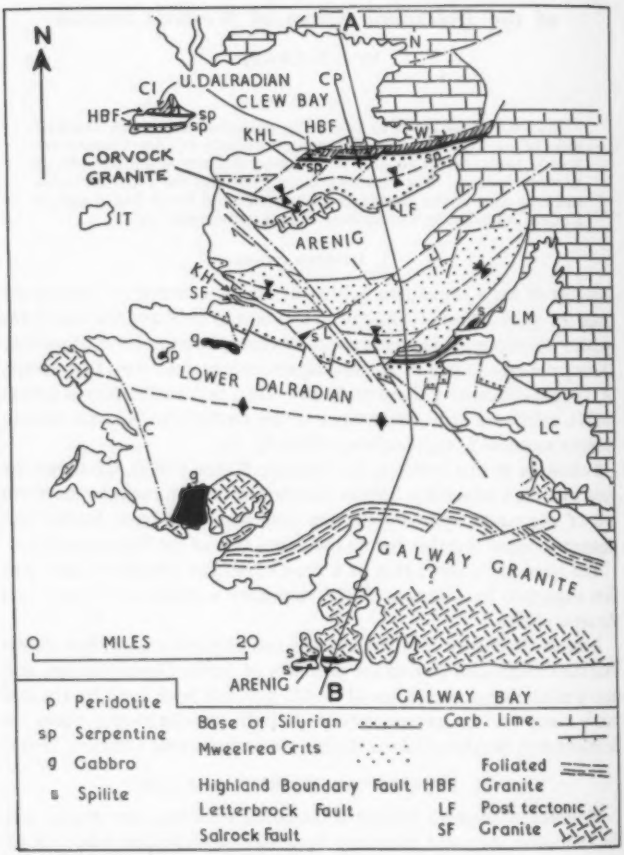
Followers of this concept, for example Kilroe (1907), advanced the view that the Connemara Schists and the Ordovician rocks to the north are of the same approximate age and that they were folded and metamorphosed together before the deposition of the Silurian rocks.

The alternative view, that of a pre-Ordovician metamorphism, was first suggested by Callaway (1887) and later amplified by Wager and Andrew (1930).

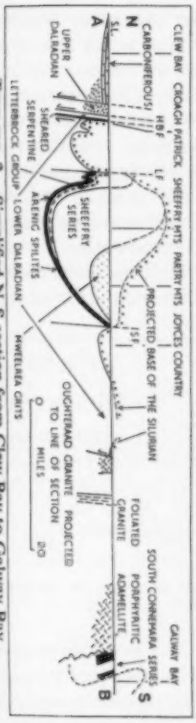
Recently Professor Shackleton (personal communication) has shown that the Connemara Schists are probably of Lower Dalradian age, and also a minimum absolute age of − 475 m.y. has been fixed for the last main event in the metamorphism of both the Dalradian rocks of Scotland and Western Ireland (Giletti, Moorbath, and Lambert, 1961).

II. Evidence for Younger Age Limit

Nowhere in Western Ireland is the contact between the Arenig and the Dalradian seen and, therefore, there can be no precise definition of its nature. The lowest visible part of the Arenig succession consists of a group of spilites, cherts and black slates exposed south of Lough Nafooey and to the south-west of Leenane (Text-figs. 1 and 2). *Didymograptus extensus* is recorded from black slates in these rocks near Bencraff (Carruthers and Muff, 1909). Stanton (1960) has also recorded a fauna indicative of the *D. extensus* zone about 5,000 feet from the top of the Sheeffry Series south of Lough Nahaltora. Below the Sheeffry Series, on the northern limb of the Mweelrea syncline (Text-figs 2 and 3) is a sequence of slide conglomerates, turbidites, and black



TEXT-FIG. 1.—Simplified geological map of West Connaught. Key to abbreviations :—C—Clifden ; CP—Croagh Patrick ; KH—Killary Harbour ; KHL—Kilgeever Hill ; L—Leenane ; LB—Louisburgh ; LC—Lough Corrib ; LM—Lough Mask ; N—Newport ; O—Oughteraad ; P—Partry Mts. ; S—Sheeffry Mts. ; W—Westport.



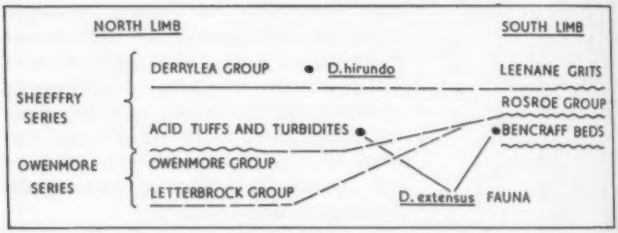
TEXT-FIG. 2.—Simplified N–S section from Clew Bay to Galway Bay.

slates which the writer would like to term the Owenmore Series (Dewey, 1960). The base of the series is not seen and the lowest part of the succession consists of a Group of boulder slides (Letterbrock Group) with a northerly provenance. These strata lie below the slates in which Stanton recorded a *D. extensus* fauna. The Letterbrock Group contains pebbles of fresh spilite and jasper which are almost certainly derived from Lower Arenig spilites, from which the conclusion must be drawn that the Letterbrock Group belongs to the *D. extensus* zone. In addition to the jasper and spilite and forming the largest proportion of the boulders are well rounded boulders of pink and white quartzite and a smaller amount of schist. Contemporary palaeogeography can be fairly accurately reconstructed and appears to be that of an east–west trough bounded to the north by a fault scarp with uplands composed of metamorphic Dalradian rocks to the north.

The upper part of the Sheeffry Series (Derrylea Group) on the northern limb of the Mweelrea syncline (Text-fig. 3) consists of turbidites derived both from the east and south with a fair amount of detrital staurolite. Small rounded schists and quartzite pebbles are occasionally present in the turbidites derived from the south which, together with the staurolite, were probably derived from the Lower Dalradian rocks of Connemara.

Immediately north of the Highland Boundary Fault, on the south shores of Clew Bay and on the northern part of Clare Island there occur Upper Dalradian greywackes, graded micro-conglomerates, cherts, limestones, and pillow lavas. Wherever Dalradian rocks are seen south of the Highland Boundary Fault, however, they are higher grade Lower Dalradian. Between Clew Bay and Kilgeever Hill, about 4 miles west of Crogh Patrick, a very instructive section can be seen (Text-fig. 4). Kilgeever Hill itself is composed of the Croagh Patrick quartzite with a basal littoral boulder conglomerate. A narrow belt of serpentine extends E.–W. along the foot of Kilgeever Hill with another belt of serpentine about 30 yards to the north. Between the two serpentine belts and immediately to the north of the northernmost are two thin strips of ground occupied by finely banded amphibolites

and garnet-muscovite schists, with the Upper Dalradian Westport Grits faulted against them to the north. The fault which throws the Westport Grits against the amphibolites and schists has no associated serpentine screen as do the two faults to the south. Clearly there must have been a considerable downthrow to the north on the northern fault to bring chlorite-grade Upper Dalradian rocks down against garnet-grade Lower Dalradian rocks. George (1960) has already suggested that the Highland Boundary Fault must have had an initial northerly downthrow of a considerable magnitude and this is merely confirmatory evidence. While this initial movement must be pre-Silurian it is also probably pre-Arenig since in North Connemara the Upper Dalradian is almost certainly absent between the Arenig spilites and the Connemara



TEXT-FIG. 3.—Comparison of the Arenig rocks on the northern and southern limbs of the Mweelrea syncline.

schists. This involves the pre-Arenig erosion of a considerable cover of Upper Dalradian rocks to expose sillimanite and staurolite grade Lower Dalradian. This faulting and erosion must post-date the metamorphism of the Dalradian and pre-date the deposition of the Arenig strata.

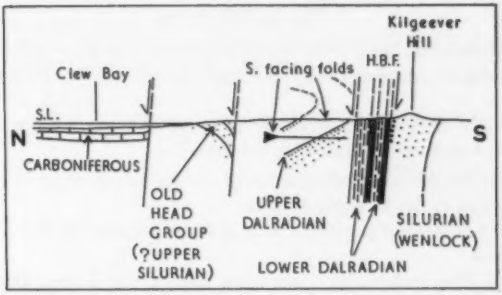
Recent work by the writer (Dewey, 1960) has shown that the unconformable relationship between the Arenig rocks and the Upper Llandovery on the southern limb of the Croagh Patrick syncline (Text-fig. 2) is the result of Ordovician faulting and a relatively small amount of (?) Taconic crustal shortening. There is no pre-Silurian metamorphism of the Ordovician rocks ; the Silurian rocks, in fact, acquired a grade of metamorphism approaching the chlorite/biotite isograd (Anderson, 1960) during the main (?) Erian metamorphism.

The Ordovician rocks of the Mweelrea syncline show a progressive southward decrease in metamorphic grade, from chlorite-slates and greywackes on the steep northern limb in Murrisk to virtually unmetamorphosed boulder conglomerates and greywackes (McKerrow and Campbell, 1960) on the southern limb in Joyces Country. Since the Ordovician and Silurian rocks acquired their metamorphism together, the Dalradian metamorphism is almost certainly pre-Arenig because

south of Killary Harbour there is an unconformity and a metamorphic break between the Connemara Schists (sillimanite-grade) and the Upper Llandovery.

South of Killary Harbour the Upper Llandovery rests upon the Connemara Schists to the exclusion of the Ordovician succession. This is mainly due to an original southward thinning against a rising Ordovician genaticline in Connemara, but Taconic tilting and faulting may be partly responsible.

Arenig rocks occur on the south side of the Galway Granite where McKie and Burke (1955) have described a thick succession of spilites and greywackes. These rocks were clearly folded and cleaved before



TEXT-FIG. 4.—N.–S. section from Clew Bay to Kilgeever Hill.

the intrusion of the main post-tectonic facies of the Galway Granite. McKie and Burke argue that the emplacement of the main Galway Granite is Taconic since the Oughteraad granite to the north, which is later than the main Galway granite, is overlain unconformably by Valentian rocks. While it is fairly certain that the Oughteraad granite is pre-Valentian, the main Galway granite has an age of − 365 m.y. (Giletti, Moorbath, and Lambert, 1961) which, according to the Revised Geological Time-Scale (Holmes, 1959) is a Devonian age.

It is probable, furthermore, that the Oughteraad granite is even older than Taconic since both the Mweelrea Grits and the Leenane Grits contain large boulders of a porphyritic granite strikingly similar to the Oughteraad granite.

The folding of the Arenig South Connemara Series probably took place during the main Erian phase and the post-tectonic facies of the Galway Granite may well be of the same age as the Corvock Granite (Text-fig. 1) north of Killary Harbour, the latter granite intruding metamorphosed Ordovician and Silurian rocks.

III. Evidence for an Older Age Limit

The latest important events in the metamorphism of the Dalradian rocks of Scotland and Ireland have been shown (Giletti, Moorbath, and Lambert) to have taken place at approximately − 475 m.y. and thus the conclusion of progressive regional metamorphism of the Dalradian appears to have been a broadly synchronous event.

The Leny Limestone, an integral part of the Scottish Dalradian succession (Stone, 1957), contains *Pagetia* (Pringle, 1940) and, therefore, appears to be of lower Middle Cambrian age. The recumbent folding and metamorphism of the Dalradian therefore took place after lower Middle Cambrian times.

IV. Conclusions

It is considered likely that the metamorphic date of − 475 m.y. for the Dalradian metamorphism is Upper Cambrian or Tremadocian since :—

(1) The metamorphism is believed to pre-date the *D. extensus* zone of the Arenig on the following evidence :—

(*a*) Boulders of Dalradian quartzite and schist occur in conglomerates in the *D. extensus* zone.

(*b*) Staurolite is present as a detrital mineral in the Sheeffry Series.

(*c*) The Ordovician rests upon high grade Lower Dalradian rocks south of the Highland Boundary Fault suggesting the pre-Arenig removal of a considerable thickness of Dalradian rocks following an initial northerly downthrow.

(*d*) There is an important metamorphic break between the Lower Dalradian and Silurian, and the Ordovician and Silurian rocks acquired their metamorphism together.

(*e*) The Oughteraad granite cuts regionally metamorphosed Dalradian rocks, and boulders of a porphyry, believed to be the Oughteraad granite, are present in the Arenig.

(2) The metamorphism is post-lower Middle Cambrian since the Leny Limestone is an integral part of the regionally metamorphosed Dalradian succession.

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Buried Valleys of the Upper Rheidol (Cardiganshire)

By T. D. Adams

Abstract

Four Boulder Clay-infilled channels of the upper Afon Rheidol (Cardiganshire) are described as relict river valleys of a progressive localized diversion by glacial interference.

The district around Ponterwyd, Cardiganshire, forms part of a dissected plateau lying between 1,000 and 1,200 feet, modified by localized glacial deposition and later fluvial erosion. Drainage is dominated by the main north to south flowing Afon Rheidol. South of the village, the Rheidol lies within a deeply dissected gorge, its sides rising 300 feet to the adjacent plateau segments. Within this river section, two westwardly curved "incised meanders" are seen. They have been considered to be the result of glacial diversion (Challinor, 1933).

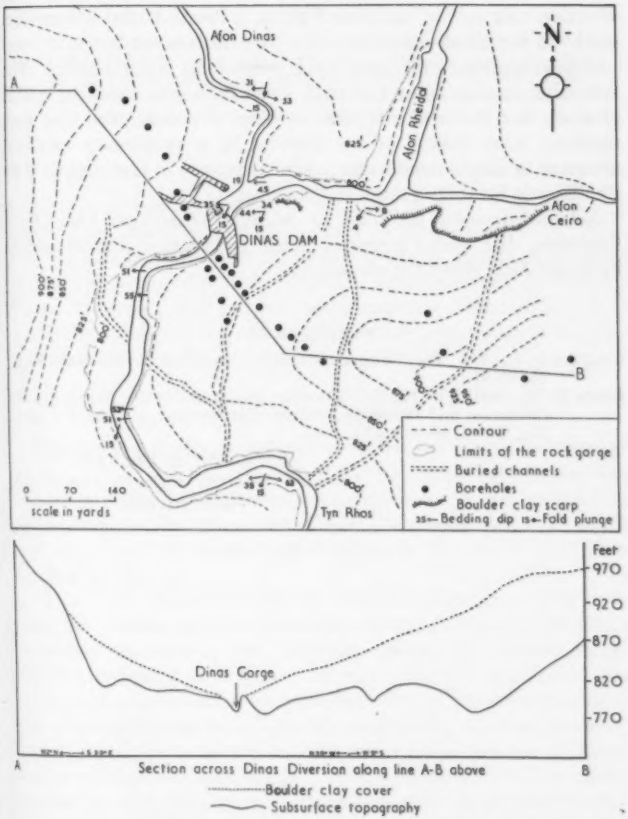
North of Ponterwyd, the cross section of the main valley becomes assymmetrical, with the steep-sided plateau remnant known as Mynydd Ponterwyd lying to the west, and the Brynglas foothills gradually rising to the east. A westward diversion similar to those south of the village is found at Dinas (Grid ref. O.S. 1-in. sheet 127 E745, N820). This has recently been studied in connection with the construction of a dam and reservoir.

The plan and scale of the Dinas Diversion can be seen on the accompanying map. To the west, the river cuts through a steep-sided rock gorge, commencing at the Dinas Rheidol confluence, ending below Tyn Rhos. The gorge deepens southwards, the thalweg becoming very broken by a series of minor waterfalls and rapids.

The solid rocks at the diversion belong to the Upper Ordovician Brynglas group and the Eisteddfa and Lower Rheidol Groups of the Llandoverian (O. T. Jones, 1909). They form a highly cleaved succession of mudstones and thick well-jointed grits, folded asymmetrically on NNE-SSW pitching fold axes. Near the western portion of the diversion, the river has eroded along the strike and cleavage. To the north and south downcut has been across the structural trend.

These basement rocks are for the most part overlain by a thick mantle of variable glacial and fluvioglacial material. Exposure of the clays is extremely poor, most of the deposit being covered by a dense moorland vegetation. Nevertheless two distinct clay types are observed. Firstly there is a dominant unsorted till, consisting of a fine grained, very cohesive yellow-brown to grey clay, within which there is an indiscriminate distribution of mudstone and grit pebbles (max. length elongation 6 inches). Secondly within the till there are very subordinate pockets and lenses of distinctly graded clay with included bands of

flat-lying, water-worn and size-sorted pebbles (max. sizes ½ in. × ¼ in. × 1 in.). Throughout the deposit as a whole, there is a general increase in pebble size and content eastwards.



TEXT-FIG. 1.—Buried Valleys of the Upper Rheidol.

The relation between drift and solid rock is well exposed to the north of the diversion, where the east–west flowing Rheidol has cut below the contact. Here also the drift-solid contact can be observed in plan, as "islands" of rock protruding from beneath the Boulder Clay cover. Two drift-infilled rock hollows are visible lying on the eastern and western flanks of a major breached anticline. Boreholes driven along a

critical section across the diversion area enabled further plotting of subsurface topography, whilst stereoscopic study of selected aerial photographs gave the exact locations of the infilled channel courses.

These studies show that three infilled channels are to be found in the diversion area east of the present gorge, a fourth buried channel lies partly on the hill slopes to the west. The easternmost buried channel continues southward the north-south valley form of the Rheidol. The remaining channels repeat this trend. From the deposition of the infilled channels and the nature of their infilling, it is concluded that they represent relict valleys of the Rheidol in a progressive localized diversion by glacial interference, a similar process to that suggested by Challinor in his studies south of the village.

Acknowledgments are due to Messrs. Whatlings, Ltd., Civil Engineers, who kindly permitted access to borehole records and to Professor Alan Wood for his guidance.

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On Garnet Coronites from Koraput, Orissa

By Mihir K. Bose

Abstract

Development of garnet coronas in a small meta-mafite body, forming a tilted dome within the parametamorphics, is reported and the petrography of the coronites is described. Garent coronation marks a stage in the metamorphic and textural evolution of the rocks. The origin of coronas and related features is discussed.

Introduction

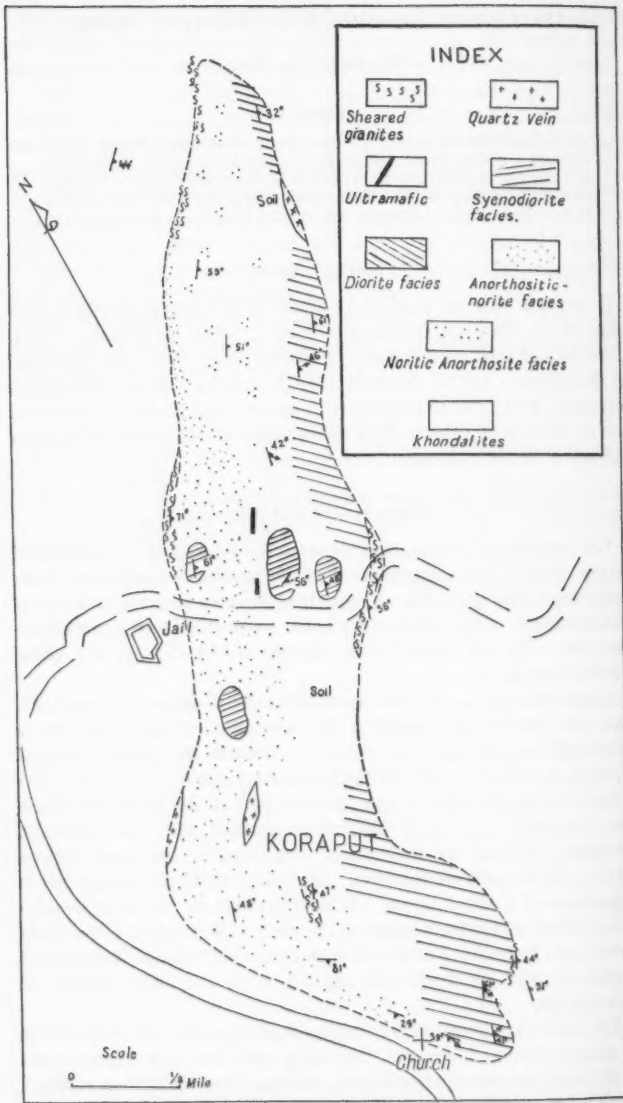
Coronites or reaction rims are of great interest to petrologists as they seem to be chemical reactions written in stone (Sederholm, 1916). Coronas and associated features of similar nature have been described by Buddington (1939), Gjelsvik (1952), Murthy (1958), and others. Reactions giving rise to coronas with garnet have played an important role in the metamorphism of the gabbroic-anorthosite body of Koraput (82° 43′ : 18° 49′), Orissa.

Geological Setting

The coronites are common in the western marginal part and frequent in the central and eastern parts of the gabbroic-anorthosite body occurring within garnetiferous-quartz-sillimanite gneisses and schists (Khondalites). The body is of lenticular shape tilted to the east-south-east and with the longest axis running N.N.E.–S.S.W. for about 1 mile (Text-fig. 1).

Linear and planar structures are normally ill-defined in the body. Near the western part imperfect platy structure, defined by felspar, is developed occasionally and foliation, defined by distinct mineral banding, is noticed locally along the eastern margin of the body. The foliation along the eastern and southern part of the body dips below the Khondalite and is, in general, concordant with the boundary. Lineation, defined by mafics and occasionally developed felspar tablets, pitches with the foliation. The overall structure of the body is suggestive of a tilted dome. Felspar crystals in the south-western margin pitch vertically, suggesting the steep walled nature of the body in this part. Imperfect fracture cleavage across the foliation is developed almost all over the body and appears to be the latest imprint of deformation.

The body may show different mineral paragenesis and proportions in different parts. In general, the rocks along the easter (upper) part of the body are richer in hornblende and may be designated as rocks of dioritic facies. The body in the central portion is relatively richer in



TEXT-FIG. 1.—Map showing the spatial distribution of different lithologic facies within the gabbro-anorthosite body of Koraput, Orissa.

plagioclase with a low percentage of mafics, i.e., anorthositic in character, while along the western and south-western (lower) part both plagioclase and pyroxene are abundant, giving a gabbroic stamp to the rocks. Pink garnet is developed almost uniformly in the body.

The body is also associated with thin sheets of younger sheared granite, occurring either within the body or between the body and the wall rock, but striking parallel to the country rocks. Coronites are, however, not spatially related to the quartzo-felspathic intrusives.

Petrography of the Coronites

The garnet coronite in its most typical development is found in the north-western margin of the body. Here it occurs as a dark band with laths of dark bluish-grey felspar, pink garnet, and darker mafics. But corona structure is also found in rocks of varying character, as noted previously. Thus the coronites may be noritic rocks (resembling basic charnockites) with greyish felspar and mafics, or mesocratic or even leucocratic types, composed dominantly of greyish or bronzy felspar.

Plagioclase is by far the most abundant of the primary constituents and the textures of the rocks are dominated by the typical lath-shaped form of this mineral. The interstices between the laths are filled by other minerals, including plagioclase itself. The coarser plagioclase individuals may have a dimensional preferred orientation parallel to the trace of the foliation. This essential primary feature is modified by garnet coronas developed between plagioclase and coloured minerals. There is practically no variation in the composition of the plagioclase, an intermediate andesine, from the margin to the centre of the intrusive body. The interstitial grains are normally acid andesine. Both coarse and interstitial grains, particularly the latter, show faint zoning.

The plagioclase grains, particularly the coarser individuals, are charged with dense schiller inclusions oriented dominantly on (010) and less frequently on (001) (cf. Barth, 1930). The inclusions are both opaque and non-opaque. The opaque inclusions have been identified as haematite and magnetite from study in reflected light and from X-ray powder photographs. Of the non-opaque inclusions biotite and occasionally amphibole are readily identifiable under very high magnification. Besides these there are abundant bright green rods with positive elongation (having closest resemblance to neighbouring pyroxene), and some flaky colourless mineral with inclined ends. Opaque rods are found closely associated with the greenish inclusions. Biotite and the colourless flaky mineral may enclose the greenish rods. It is seen that the inclusions are formed of minerals which show a definite order of crystallization among themselves. Similar oriented

inclusions in plagioclase have been noted as iron ores (MacGregor, 1931), rutile (Mawdsley, 1927), magnetite (Huang, 1955), pyroxene (Buddington, 1939 ; Barth, 1930) spinel parallel to (010) (Friedman, 1955), mica-like mineral (Gjelsvik, 1952), and magnetite, ilmenite, hercynite, pyroxene (Osborne, 1959, personal communication). Marginal parts and centres of imperfect zonation in plagioclase are normally free of inclusions. The clear fringe follows the corona boundary with garnet and appears to be related to corona formation.

The coarser individuals of plagioclase are almost always twinned on the combined Carlsbad-albite law. The smaller individuals are either untwinned or twinned on the albite law. The mineral, particularly the coarser individuals, gives evidence of post-crystalline deformation in the form of crystal bending, twin bending, obliteration of twin bands, grain faulting, and rotation. Different sets of apparently dissimilar twinning on either side of a fracture may suggest post-deformational twinning but orientation on the universal stage proves them to be rotated twinned sub-individuals.

Pyroxene, the other important essential mineral is normally an orthopyroxene and interstitial in its development. Occasionally it is coarse enough to include ores, apatite, and plagioclase. A parting is developed parallel to (100), and very fine lamellae oriented parallel to (100) are occasionally noticed, but their extreme fineness limits specific determination. Rectangular schiller inclusions of opaque ores are developed along the cleavage. The mineral is pale green and occasionally faintly pleochroic to tints of pale pink. β refractive index (1·698–1·702) and optic axial angle ($2V\alpha = 80°–86°$) suggest slight variation in composition around En 63. Occasionally some of the grains are imperfectly zoned.

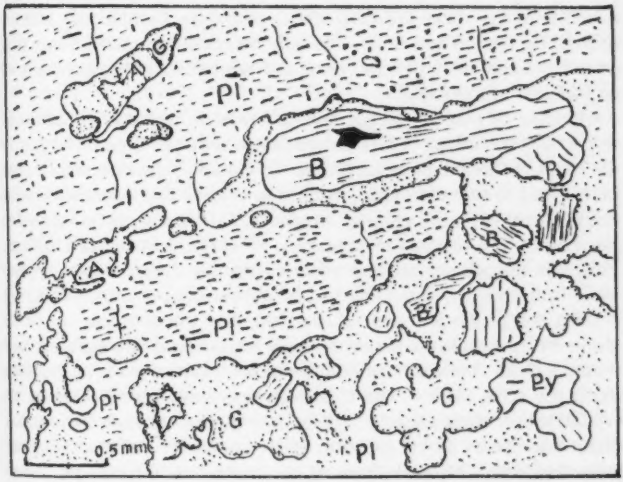
Amphibole is not common in the coronites but when present it is dominant over pyroxene, associated with the latter, and surrounded by a complete rim of garnet. The optical characters are as follows : X = lemon yellow, Y = greenish brown, Z = brown, $Z \wedge c = 14°$, $\beta = 1·684$, $2V\alpha = 83\frac{1}{2}°$. Bluish green amphibole is also developed as narrow irregular rims round pyroxene and ores.

Biotite is very frequent in the coronites, often in coarse blades without any strong orientation. In most cases the mineral shows a core of opaque ore and outer shell of garnet. The mineral shows replacement relations to hypersthene and amphibole occasionally. Pleochroism is almost always golden yellow to dark brown, $\beta = 1·660–1·683$. In association with brown hornblende, its absorption colours are pale yellow to dark reddish brown with $\beta = 1·645$.

Microperthite, developed in some of the coronites, embays other primary minerals particularly plagioclase. The mineral is typically interstitial in development when occurring in smaller grains. *Ores* and

apatite are frequent amongst the accessories. Both crystallized prior to the primary mafics but later than coarse plagioclase. Textural relations are always suggestive of apatite preceeding ores in crystallization. Apatite may show rows of dusty inclusions, resembling clouding of plagioclase. A garnet shell often develops directly between ore and plagioclase.

Garnet is the only important secondary mineral and is normally developed between mafics and plagioclase. Very often the mineral is



TEXT-FIG. 2.—Sketch showing the texture of coronite. A camera lucida sketch showing biotite (B) with an inner core of ore (black) and outer rim of garnet (G). Garnet partly enclosing apatite (A). Plagioclase (Pl) is clouded.

idioblastic towards plagioclase and xenoblastic in the inner contact (Text-fig. 2). The garnet shell often encloses unreplaced plagioclase in optical continuity with coarser individuals. Occasionally the nature of the two resembles a symplectic intergrowth when replacement has advanced further, and the idiomorphic habit of garnet in this stage is conspicuous. Such typical coronites may grade to rocks where instead of a continuous garnet shell, individual garnet crystals are developed still at their original contacts, or within the interspaces between plagioclase, or even as disseminated grains within plagioclase situated close to the imperfect shell. The most conspicuous corona development

occurs around biotite associated occasionally with a film of secondary granular felspar development in between (cf. Buddington, 1939 ; Chatterjee, 1959). Inclusions of opaque and bright green rods are often retained in the garnet with the same orientation as in the adjoining clouded plagioclase.

Petrogenetic Considerations

The gabbroic body of Koraput has been subjected to post-consolidation alteration and it is often difficult to discriminate the primary and secondary features in the rocks. The important primary minerals in the rocks are plagioclase, pyroxene, hornblende, biotite, microcline, apatite and ores. The only common secondary mineral is garnet ; occasionally felspar, ores, and doubtfully hornblende are to be included in this group.

Clouding of plagioclase has been interpreted variously by different workers. MacGregor (1931) demonstrated that clouding of felspar is the result of thermal metamorphism. In Koraput, the clouding cannot be ascribed to thermal metamorphism as corona development is quite independent of local granitic invasions. Other workers consider clouding to be a regional metamorphic effect (Poldervaart and Gilkey, 1954). In the present case the minerals forming the inclusions appear to be iron ores, pyroxene, biotite, amphibole, that is those occurring outside the plagioclase individuals. At least in one case a coarse biotite blade outside the felspar grains is seen to send into a plagioclase crystal a chain of fine rods of biotite which continue for a long distance along the (010) plane. The inclusions may have been produced by crystallization of the existing pore fluids migrating through the internal phase boundaries or by diffusion of the mineral constituents during metamorphism. Minerals forming the inclusions like ore, pyroxene, and biotite are spaced widely in the order of crystallization. Besides, inclusion of opaque rods by pyroxene rods, and of pyroxene rods by biotite rods, suggests a sequence in the mineral formation causing clouding, which tallies with the order of primary crystallization. Clear patches at the centres of imperfect zonations may be due to internal phase boundaries (Poldervaart and Gilkey 1954). The clear fringes of plagioclase individuals follow the corona boundaries and appear to have formed by later diffusion of the constituents during coronation. Besides continuity of oriented inclusions within garnet, the presence of unreplaced plagioclase within garnet coronas, suggest coronation to be later than clouding (Poldervaart, 1959 ; personal communication).

Metamorphic reactions giving rise to garnet coronas have played an important role in the metamorphism of gabbroic rocks of Koraput and in this respect the metamorphism is of a type different from that commonly observed. The garnet coronas have developed outwards

towards plagioclase and the inner contact with mafics represents the nature of the primary boundary.

There is some diversity of opinion as to the origin of coronas, and summaries of the various views are given by Sedderholm (1916), Buddington (1939), and Shand (1945). Theories regarding the origin of coronas may be grouped as (*a*) late magmatic (including deuteric, autometamorphic), and (*b*) metamorphic (usually regional, occasionally contact, metamorphic).

For the formation of the coronas in the Koraput gabbros, the author definitely rejects the first possibility. In the case of reaction by residual magma the growth of corona minerals should be independent of adjacent minerals which is not the case, since coronas are developed between plagioclase and coloured minerals. Autometamorphic garnet coronas have, however, been described by Sutton and Watson (1951). There is also practically no evidence in favour of the other possibility here, a deuteric origin, the corona mineral being a typical anhydrous one. Gilson *et al.* (1928), have described coronas formed by the action of late magmatic solutions on pyrogenetic minerals. But in the present area pegmatites are not developed, indicating lack of late magmatic residual fluids, rich in volatiles. Abundance of apatite in coronites has been interpreted as a clue to the role of hyperfusibles in coronation. In the present case apatite does not represent a late magmatic product and there is no positive correlation between the amount of apatite and the degree of coronation or amount of garnet developed. It is evident that lack of water and other volatiles is an important condition for coronation. Sufficient water favours reactions outside the mineral individuals destroying the corona relation. Slight excess of water will cause recrystallization of plagioclase, which has not been the case with the present coronites or most of the coronites so far described. In a recent contribution Wentorf (1956) has shown that development of garnet from hornblende requires removal of water from the system. The present coronites and related rocks show gradual development of garnet in the interstices between plagioclase or along the (010) planes of plagioclase or as disseminated grains within plagioclase, close to the corona rim. Garnet development close to mafics may be noted with a change in the primary fabric. Ultimately garnetiferous plagioclase granulites are obtained, the mineral paragenesis of which is the same as that of the coronites. It seems most probable that coronites mark a particular stage in the metamorphic and textural evolution of the gabbroic rocks. Garnet coronation marks the advent of a metamorphic episode and not the closing of a magmatic history. The essential minerals stable during the metamorphism of the rocks are plagioclase, garnet, orthopyroxene, and potash felspar, and the mineral paragenesis is comparable with that in the granulite facies.

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DEPARTMENT OF GEOLOGY,
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An Unusual Gabbroic Intrusion from Northern Rhodesia

By A. R. Newton

Abstract

Gabbro is intruded near the contact of gneiss and schist groups in the Southern Province of Northern Rhodesia. The petrography of normal and altered rocks is described. The rocks may have originated as an intrusion into an original reversed drag fold which it has emphasized and distorted. Intrusion took place at a late stage of regional metamorphism and folding.

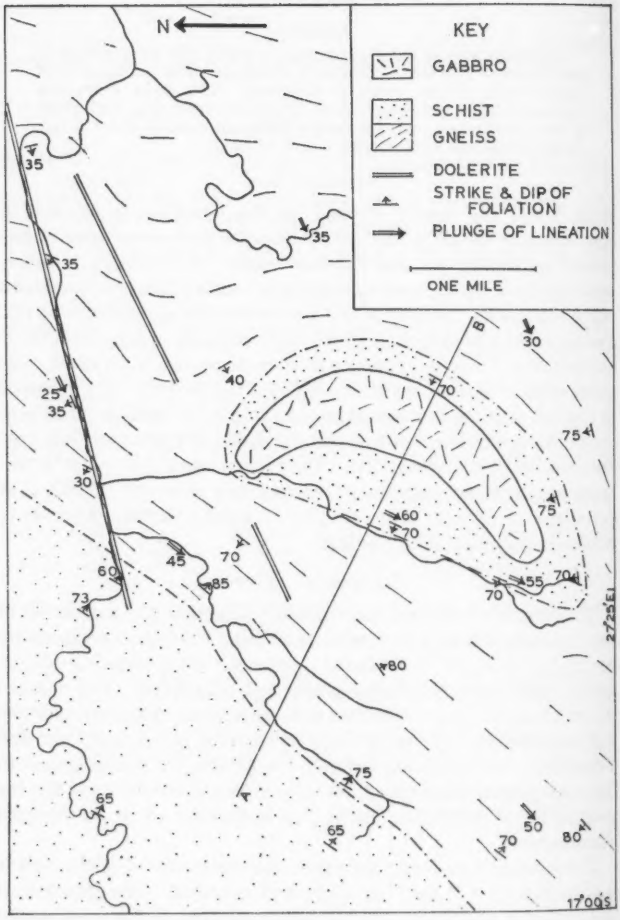
Introduction

Kala Hill, the feature formed by the intrusion, is situated at approximately 16° 59′ S. and 27° 26′ E. in the escarpment zone on the north-west side of the mid-Zambezi valley. The country is deeply dissected by a young and vigorous river system draining south-east into Lake Kariba, but even in this area of considerable relief, Kala Hill stands out as a prominent feature. It is crescentic in plan (Text-fig. 1) and just over 2 miles long, with a steep eastern slope and a much more gentle western one on the concave side of the crescent. The existence of the hill, and the fact that it consists of a basic intrusion, has been known for some time, but as far as the author is aware there has been no prior detailed investigation. The present work was part of a programme of regional mapping in the area on a scale of 1:50,000. The permission of Director of the Northern Rhodesia Geological Survey to publish is gratefully acknowledged.

Geological Setting

The intrusion is situated approximately 2 miles east of the contact of two metamorphic rock groups of contrasting lithology, a gneiss group consisting mainly of well-foliated quartz-microcline-oligoclase-biotite-gneiss, with some hornblende-gneiss and amphibolite, and a schist group, of quartz-muscovite-schist with subordinate muscovite-quartzite and amphibolite. The schists contain sporadic garnet, staurolite and sillimanite. At the contact between the groups, the schists appear to dip steeply beneath the gneiss, but from evidence elsewhere and also the orientation of local drag folds, this is thought to be an inverted relationship.

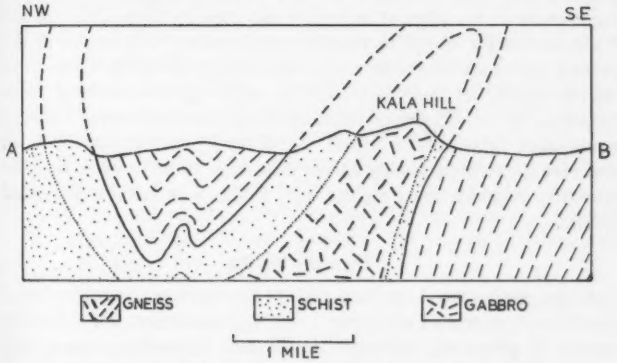
The igneous body outcrops within the main area of gneiss, but is surrounded by a "rim" of schist and quartzite. The quartzite is hematite-rich and similar to the basal members of the schist group adjacent to their contact with the gneiss. The structural interpretation is shown in the section (Text-fig. 2), from which it can be seen that the intrusion forms the core of a large-scale drag-fold-type structure overturned to the east; the map shows that the fold is doubly-plunging with a curved axial trace.



TEXT-FIG. 1.—Sketch-map of the Kala Hill area.

Petrography and Classification

Specimens taken across the centre of the body show that the rocks are composed of almost equal proportions of the three main minerals—augite, hypersthene, and labradorite. Augite is slightly in excess of hypersthene in two specimens, while this relationship is reversed in a third. The total range is augite 47·3 to 61·2 per cent, hypersthene 52·7 to 38·8 per cent (of total pyroxene). These rocks, therefore, fall in the middle of the gabbro-norite series and, following Hatch, Wells, and Wells (1949, p. 277), the rocks with hypersthene in excess of augite are augite-norites, while those with predominant augite are hypersthene-gabbros. As regards colour index, it has been shown by Ellis (1948) that the natural upper limit of gabbros is about 70. In a recent paper



Text-fig. 2.—Diagrammatic Section through Kala Hill.

by Drysdall and Stillman (in press), this has been incorporated in a general classification, and is followed here in preference to the limit of 90 used by Hatch, Wells, and Wells (1949, p. 277) and 95 by Johannsen (1939, p. 143). A minor modification by Drysdall and Stillman of Ellis' scheme is that his term gabbro-pyroxenite for rocks with a colour index more than 70 is replaced by melagabbro. The colour index of the Kala Hill rocks ranges from 65 to 72, and they, therefore, extend from the gabbro just into the melagabbro range. In thin-section (GS. 2933, 2935, 2952),[1] the feldspars show the disorientated laths typical of ophitic texture, but instead of large, enclosing augite crystals there is subhedral augite and hypersthene, and it would appear that all three crystallized more or less together. The texture is, therefore, described as subophitic.

[1] Numbers in brackets refer to thin-sections in the collection of the Northern Rhodesian Geological Survey.

Clinopyroxene is colourless augite with $c \wedge Z$ approximately 49°, and often shows many lines of tiny inclusions parallel to (001), giving a " herringbone " structure in sections parallel to (010) that are twinned on (100). The direction of these lines of inclusions is appropriate to one of the sets of exsolution lamellae which may be developed in augite (Poldervaart and Hess, 1951, p. 483), and some fine lamellae are actually visible : these extinguish at a small angle and are thought to be pigeonite. The tiny inclusions are probably also exsolution features. There is some slight marginal alteration to pale-green amphibole. Hypersthene shows typical pleochroism and tends to be more euhedral than augite, often occurring as large columnar crystals with prominent transverse cracks. Some crystals show very narrow, spindle-shaped polysynthetic twins parallel to *c*. Alteration to a pale-green amphibole commences at the edge of the grain and extends along cleavage and fractures until a complete pseudomorph results. All stages in this process can be seen in the same thin-section (e.g. GS 2933). Labradorite occurs mainly as well formed laths showing characteristic albite twinning. It is fresh, but contains clouds of dusty inclusions which, in many cases, fill the entire crystal. In others, a narrow clear rim is left and rare gradational zoning shows these rims to be more sodic. The accessory minerals are irregular iron oxides with a little associated biotite.

Altered Rocks

At the northern and southern ends of the outcrop, the rocks show a considerable degree of alteration : in the least-altered, the subophitic texture is preserved, although augite and hypersthene have been replaced by hornblende. In other cases, the feldspar is partly granulated and all trace of igneous texture destroyed. Finally, some of the hornblende-plagioclase rocks are faintly foliated. In all these cases, the feldspar remains labradorite. Hornblende, which replaces both pyroxenes as subhedral shreds and laths, has $c \wedge Z$ about 27°, and a large 2V ; its pleochroism formula is as follows : X = straw colour, Y = olive-green, Z = pale green, $Y > Z > X$. Accessory minerals include quartz (a product of the pyroxene → amphibole reactions), scattered iron oxides with associated flakes of biotite, granules of sphene and rare apatite.

Emplacement

In cross-section the fold in which the gabbro is emplaced appears to be of simple drag fold type and, if interpreted on this basis, would imply that the whole sequence is overturned and that the gneiss should be stratigraphically above the schist. The following factors make this interpretation unlikely :—

(1) The "drag fold" plunges steeply at each end, and its strike length is limited to that of the intrusion.
(2) No other drag folds with this orientation have been observed.
(3) Small-scale drag folds of more normal type in the main schist sequence imply that the schists are stratigraphically above the gneiss.
(4) An unknown amount of the deformation must be due to the intrusion itself and, unless this can be allowed for, the deformation due to tectonism alone cannot be accurately assessed.

Thus it seems that another explanation must be sought for the orientation of the fold, and, while no completely satisfactory hypothesis has been evolved, it is thought that a combination of two factors might have produced the observed result.

(1) On a regional scale, the folding is broadly anticlinorial in the east and synclinorial in the west, with Kala Hill on the intermediate complex limb. Thus the original fold might be a reverse drag fold due to uplift of the anticlinorium, of the kind described by Lambert on a much smaller scale from Morar (1958, p. 182–3).
(2) This original fold was probably emphasized and distorted by the intrusion, the upward pressure of the mass causing an arching of the axis, and possibly the weight of the intrusion causing a sagging of the isoclinal envelope to produce the crescentic curvature of the axial plane.

This naturally implies that the intrusion took place at a time when the country-rocks were still hot, and possibly plastic. Evidence supporting this is :—

(1) The lack of a metamorphic aureole round the intrusion ; (2) the lack of a chilled margin ; (3) the concordance of the envelope with the intrusion.

The last also implies that the final phase of intrusion had been reached and that the gabbro had insufficient energy left to penetrate the envelope. That the folding movements were also in their final stages is shown by the slight degree of alteration of the gabbro, which suggests no more than deuteric action or mild dynamometamorphism.

The gabbro shows no affinity to the nearby dolerites, which are post-tectonic intrusives of Karroo type.

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The Non-marine Mollusca of the Cainscross Terrace near Stroud, Gloucestershire

By N. F. Large and B. W. Sparks

Abstract

Recent collecting has revealed a typical cold mollusc fauna in the Cainscross Terrace of the Frome valley near Stroud. Cold conditions are also indicated by lithology and vertebrates. The most likely date seems to be the Irish Sea Glaciation, i.e. the main phase of the Last Glaciation.

Of the three terraces recognized by most investigators (e.g. Gardiner 1938, Tomlinson 1941) in the obsequent valley of the River Frome, the middle or Cainscross Terrace is the best developed and the most rewarding in its yield of fossils. The largest patch of this terrace gravel, excluding that at Frampton, which, although attributed by Gardiner to the Frome, is more likely to have been deposited by the Severn, is in the vicinity of Stroud. It has been exposed in two main localities, earlier at Gannicox (Grid Ref. 31/843053) and, at the present time, a mile further west at Cainscross (31/834053).

In both localities the general lithology seems to have been similar, so that the present section, noted below, will suffice as a general description of the gravel.

(3) 5 feet of roughly-bedded, but ill-graded, Jurassic gravel.

(2) 5 feet of buff loam, mostly a mixture of silt and fine sand, which has increased in thickness as the pit has been cut inwards. This loam contains small ball-like inclusions of tenacious blue-grey clay and also lenticles of fine current-bedded gravel.

(1) 10 feet or so of generally coarse, ill-graded gravel, containing lenses of sand and silt, some of which are contorted at their ends. Similar lenses of " clay " were recorded from the earlier Gannicox exposure (Lucy, 1872).

The terrace has produced both molluscs and vertebrates, the latter, which included principally *Mammuthus primigenius* (= *Elephas primigenius*) and *Coelodonta antiquitatis* (= *Rhinoceros tichorhinus*), clearly indicating cool or cold, open vegetation conditions. The molluscan evidence has been equivocal.

The earliest list of species was given by Lucy (1872) and attributed to the lenses of clay in the Gannicox pit on the authority of a paper by Jones and Witchell. No exact reference is given to this paper and, although it is said to be in the same journal as Lucy's paper, a search

through the earlier volumes has failed to reveal it. The species listed were :—

Lymnaea truncatula
Lymnaea peregra
Lymnaea auricularia
Ancylus fluviatilis
Pupilla muscorum
Cochlicopa lubrica
Helix nemoralis
Zonitoides excavatus

In the light of later collecting there are one or two peculiarities in this list, so that one wonders about the provenance of the material and the identifications. *L. auricularia* was possibly only an inflated *L. peregra*, and could the *Z. excavatus* have been the *Hygromia hispida* of later lists ? *Z. excavatus*, normally considered a calcifuge, is an oddity in this highly calcareous district, although it has recently been discovered in calcareous Pleistocene deposits (Kerney, 1959). *H. nemoralis* has not since appeared in the loam, neither has *A. fluviatilis*, usually characteristic of rather rapidly moving water, a condition not indicated by later collections.

The next list from Gannicox is fortunately on the authority of Kennard and is available in three slightly different versions : in a short note by Gardiner (1936), in a longer paper by Gardiner (1938), and in Kennard's MS notebooks which accompany his collection at the British Museum (Natural History). These contain probably his final thoughts on identification and also the numbers of shells seen, so that we reproduce below that version of the list for which we are indebted to Mr. C. P. Castell of the British Museum.

Valvata piscinalis (Müller)	1
Lymnaea truncatula (Müller)	18
Lymnaea peregra (Müller)	5
Planorbis leucostoma Millet	2
Succinea oblonga Draparnaud	1
Succinea pfeifferi Rossmässler	11
Pupilla muscorum (Linné)	174
Hygromia hispida (Linné)	210
Limax sp.	3
Pisidium sp.	2

Kennard's comments on age and climate were non-committal, although he considered that there was plenty of grassland.

Tomlinson (1941) reported almost exclusive *Pupilla muscorum* in some loam horizons at Cainscross, but does not seem to have investigated further.

Five samples recently taken from the loam at Cainscross yielded the following :—

	A	B	C	D	E	Total
Lymnaea truncatula (Müller)	1	4	—	—	1	6
Lymnaea peregra (Müller)	1	3	—	—	—	4
Succinea arenaria (Bouchard-Chantereaux)	2	—	—	—	—	2
Columella columella (Benz)	43	—	2	—	—	45
Pupilla muscorum (Linné)	399	3	41	22	—	465
Agriolimax cf. *agrestis* (Linné)	4	—	—	3	1	8
Pisidium nitidum Jenyns	—	2	—	—	—	2
Totals	450	12	43	25	2	532

Many of the specimens are very badly preserved and some of the identifications, except in sample A, must be considered tentative.

The absence of *Hygromia hispida*, the most common species in Kennard's list, is curious, but may be partly explained by local ecological variation and partly by the fact that earlier collecting would have been from the face by eye, whereas the present samples were washed down and sorted under a low-power binocular microscope. More important, however, is the discovery of *Columella columella*, which, in association with such a high percentage of *Pupilla muscorum*, is a clear indication of cold, dry, open conditions. Some of the latter species are very elongated and seem to bear to the normal form of the species the same sort of relationship that exists between *C. columella* and *C. edentula*, though without the expansion of the last whorl found in *C. columella*.

The mollusc fauna of Cainscross is now seen to agree very closely with the vertebrate fauna and with the lithology of the material. In an appendix to Gardiner's paper (1938) Sandford ascribed the elephant teeth all to *E. primigenius* and suggested a grass-steppe habitat, while the coarse, ill-sorted, ill-bedded gravels seem to indicate frost-weathering on the uplands and the transport of detritus into the valley under cold conditions. Tomlinson (1941) further commented on the fact that the gradient of the terrace is steeper than the present valley gradient, a feature to be expected of a stream with a high load/discharge ratio such as might be expected under cold conditions. She also suggested that the Cainscross Terrace expands into a fan at Eastington and correlated it with similar features elsewhere at the foot of the Cotswolds. Everything, then, points uniformly to cold conditions prevailing during the formation of the terrace.

The age of the deposit is more difficult to determine. These loess-type mollusc faunas, although usually Last Glaciation in age, are not confined to that period either here or on the Continent (Sparks, 1957). Although in fauna and lithology this deposit is extremely like some of those in the east of England e.g. at Little Chesterford, Essex (Sparks, 1955), and at Thriplow, Cambridgeshire (Sparks, 1957), it lies far to the

south-west of any previously known mollusc fauna of this type, and, in any case, both fauna and lithology can be ecological as well as time indicators. Sandford, on the basis of the elephants, suggested an older rather than a young Pleistocene date for the terrace, but both Gardiner and Tomlinson correlated the deposit with Wills' Severn Main Terrace, which has a cold vertebrate fauna and is the equivalent of the Irish Sea Ice, i.e. the main phase of the Last Glaciation. This seems to us to be the more likely correlation for, apart from the fact that most cold faunas of the Cainscross type are of this age, the gravels show none of the surface weathering nor frost-action that one would have expected had they been deposited during an earlier glaciation and hence been subjected to both interglacial and periglacial weathering.

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DEPARTMENT OF GEOGRAPHY,
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A Note on Greisening in the Bodmin Moor Granite

By C. S. Exley

(PLATE XXII)

Abstract

Some differential weathering of tors on Bodmin Moor has been found to be due to narrow zones of greisen. These zones are not related to the joint system in the granite and appear to have arisen by the alteration of feldspars in the path of escaping late-magmatic fluids.

I. Introduction

During recent field work on Bodmin Moor, it has been observed that many granite tors have rather corrugated surfaces as the result of differential erosion. A typical example which has been examined in some detail is the small tor at the southern end of Brown Gelly (Plate XXII A), about three miles north of St. Neot.

As a rule, the ridges in such cases, which are 1–2 cm. high, are due to quartz or quartz-tourmaline veins, or, more rarely, to narrow aplite veins, but a number of the ridges examined have been found to contain no such obviously resistant material. Indeed, no explanation for their presence could be discovered in the field.

II. Nature of Ridges

The granite upon which the ridges form is typical of granites from Cornwall and is deeply weathered. It contains the usual suite of minerals, viz. quartz, microperthite, normally-zoned plagioclase ranging from ca. An_{30} in the core to ca. An_{24} at the rim, biotite, muscovite, and tourmaline. There has clearly been much recrystallization for a great deal of secondary quartz and mica are present and the other minerals all show evidence of attack and replacement. Nevertheless, the medium- to coarse-grained, slightly porphyritic, granitic texture is still clearly preserved in the rocks of both troughs and ridges, the general appearance of which is identical until they are cut and polished.

1. *Veins.*—An examination of both polished and thin sections of granite taken from the position indicated on Plate XXII A reveals that the rocks forming the ridges differ from those forming the troughs in having narrow veins, never more than 0·5 mm. wide, of micaceous material. These veins traverse the rock haphazardly, serving as links between areas of altered feldspar. They tend to follow crystal boundaries but occasionally cut across otherwise undisturbed crystals of feldspar and mica. It is rare to see veins in quartz crystals, although a concentration of inclusions with a similar orientation is sometimes found in quartz, aligned with the end of a vein in a feldspar or mica crystal.

In composition the veins consist almost entirely of minute flakes of white mica, often in radiating clusters, and occasionally a little irregular quartz. Being near the ground surface, the veins are often stained brown.

2. *Altered feldspars.*—A second distinctive feature of the ridge-forming granite is the degree of alteration of the plagioclase. In the main part of the rock and in that forming the troughs it is cloudy and frequently partially altered to clay and secondary mica, especially in the cores of zoned crystals. The margins are often rather irregular, and a distinguishing feature of the secondary mica is its low birefringence. In the rocks of the ridges, on the other hand, the plagioclase is far more extensively altered to scattered flakes of a secondary mica with a high birefringence and to quartz. In many cases replacement is complete. Unlike secondary mica in the microperthite, that in the plagioclase seems to develop without any obvious reference to crystallographic directions.

Throughout the rock, in both ridge and trough modifications, the microperthite is cloudy and is partially replaced by quartz, which embays the margins, and white mica which develops along cracks and cleavages. The amount of alteration is small in general and is exactly like that seen elsewhere in Cornwall (e.g., Exley, 1959, pp. 211–2).

It appears, therefore, that the rock forming the ridges consists of zones, between 1 cm. and 2 cm. wide, of incompletely greisened granite (Plate XXII, B and C). Its resistance to what is mainly chemical erosion arises from its high proportion of mica and quartz, both of which are, of course, relatively stable in atmospheric conditions.

III. Origin of Ridges

With the exception of the widespread but incomplete alteration of potash feldspar found in some Cornish granite (Exley, op. cit.), greisens have been thought to occur usually in association with fissure systems, expecially where mineralization has taken place, and many examples from Cornwall have been described (e.g., Hill and MacAlister, 1906, p. 58 ; Hosking, 1951, p. 343). Such greisens are clearly due to the introduction of altering solutions subsequent to consolidation and fracturing of the granite. The Brown Gelly ridge-forming greisens, however, show no sign of being linked with any fissures, and their strikes differ by 10 to 20 degrees from those of the joints whose directions are most similar. Moreover, the dips of these greisens are generally some 10 degrees steeper than those of the joints.,

There can be little doubt, therefore, that the zones of greisening were established while the granite was still hot, although the fact that

micaceous veins sometimes cut across quartz crystals suggests that the crystallization of the main minerals was essentially complete. It is difficult to believe that the fluids responsible for the greisening would have worked through solid rock, from feldspar to feldspar, had a fissure system existed by which they could have travelled. Again, the difference in strike and dip suggests that there was a change in the direction of minimum pressure between the initiation of the greisen zones and the development of the joints ; it is only fair to add, however, that of the joint-directions measured over the southern half of the Bodmin Moor granite, about 20 per cent conform with the greisen-directions of Brown Gelly whose divergence there may be a local phenomenon.

In a discussion on the processes giving rise to greisens, Williams (1934, p. 349) has pointed out that these may operate more or less continuously throughout the last-magmatic, earliest-crystalline and post-jointing stages of granite-formation and that the whole trend of magmatic differentiation in granites is towards the production of liquids capable of giving greisens. Both the widespread late-magmatic alteration of feldspars and the introduction of greisening fluids through joints in Cornish rocks has been mentioned above, and the production of the ridge-forming greisens appears to represent a stage intermediate between these two extremes.

If Williams' (op. cit., p. 346) criterion for deciding on whether water or fluorine predominated in the altering fluids (viz. the presence or absence of topaz) be accepted, it is evident that the Brown Gelly fluids were mainly aqueous.

Most descriptions of Cornish greisens do not specify whether alteration has started with plagioclase or potash feldspar but the writer's experience suggests that the latter tends to be attacked first and provides a source of potash that is used later in various ways. Evidence from the granites of St. Austell (Exley, op. cit.), Tremearne and Godolphin (M. Stone, private communication) and Carnmenellis (W. G. C. Austin, private communication) indicates that commonly in Cornwall much recrystallization has taken place during those critical stages in magmatic evolution when water, silica, potash, and soda are greatly concentrated. In these circumstances it is to be expected that both the acidity and the concentration of a particular alkali would have varied from place to place and from time to time and that various kinds of auto-metasomatism would have occurred. One of these is greisening and in the cases where plagioclase has been replaced, as at Brown Gelly, the altering solutions, while acid enough to destroy the original crystals, contained sufficient potash to enable the critical ratio of $K_2O : Al_2O_3$, required for the synthesis of mica, to be reached (Gruner, 1939, 1944).

The character of the greisen zones shows that the movement of the fluids was restricted and one is compelled to suppose that weaknesses in the crystal mesh (arising from cooling or tectonic causes) favoured migration in a particular direction. Doubtless the increased porosity of altered feldspars encouraged the circulation of fluids in their vicinity and thus the greisen zone would extend, its two-dimensional nature being controlled by the direction of the original zone of weakness.

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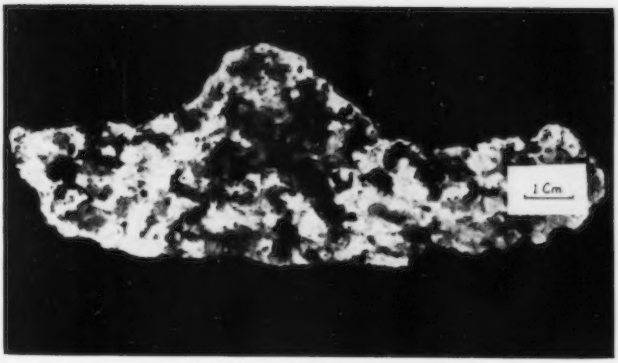
EXPLANATION OF PLATE XXII

A.—Small tor at southern end of Brown Gelly, showing ridges due to differential erosion of veins and greisened regions. Circle indicates locality from which specimens were taken.

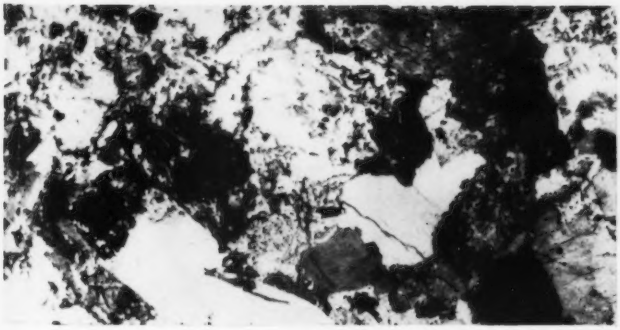
B.—Polished section cut through ridge of greisen and flanking troughs of unaltered granite. Altered feldspars and small veins are clearly visible in central part of specimen.

C.—Thin section (polarized light, × 1) of part of greisen ridge. Relatively unaltered microperthite (twinned, bottom right-hand corner ; in extinction, centre right), contrasts with replaced plagioclase (centre left ; euhedral pseudomorph, bottom left). Section shows ramification of secondary mica through unfissured rock.

A



B



C

GREISENING IN BODMIN MOOR GRANITE.

facing p. 430.

CORRESPONDENCE

ORIGIN OF ALBITE PORPHYROBLASTS

SIR,—In his letter on this subject in the May–June number of this *Magazine*, Dr. A. F. Trendall is in error in supposing that Dr. K. A. Jones (*Geol. Mag.*, **98**, p. 43) has confused his (Trendall's) view with that of Reynolds in the sentence " The origin of the porphyroblasts was considered by Trendall to be due to a soda metasomatism, possibly from a trondhjemitic magmatic source ". In my 1942 paper on albite-schists to which Dr. Jones referred, the petrochemical evidence I recorded led me to conclude that " If the process of introduction of Na and Si, reinforced by Ca . . . concomitant with driving out of K, Fe, Mg, and Al had continued, the albite-schists would eventually have attained the composition of trondhjemite ". On the basis of previously recorded experiences, both of my own and of other investigators, I suggested that trondhjemite evolved in this way might become rheomorphic. I have never, in fact, considered trondhjemite magma to be the source of the soda.

For nearly thirty years I have vigorously combated the view that granitic magma is the *cause* of feldspathization. During this time, all the evidence I have observed and recorded, as well as the evidence I have read about, has led me to think that magma of granitic composition is formed from older rocks as an *end-stage* of metasomatic change when accompanied by rising temperature.

Dr. Jones' misrepresentation of Dr. Trendall's conclusions, and Dr. Trendall's misrepresentation of my conclusions are, unfortunately, but a minor symptom of a widespread malaise of geological writings of the present time. It therefore behoves us all to keep in mind Josh Billings' conclusion that " the trouble with most folks is not so much their ignorance as their knowing so many things which ain't so ". In our enthusiasm, however, do not let us forget that it was neither Reynolds (1942), nor Trendall (1953) nor Jones (1961) who first recognized that the albite-schists under discussion result from a large-scale impregnation of sodium, and that the growth of the albite porphyroblasts was commonly post-tectonic. The honours go to Clough (1897).

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DORIS L. REYNOLDS.

3rd July, 1961.

AN OLIVINE-BEARING HORNFELS FROM SOUTH-EASTERN QUEENSLAND—A CORRECTION

SIR,—Since publication in 1959 of my paper " An Olivine-bearing Hornfels from South-Eastern Queensland " (*Geol. Mag.*, **96**, 377–384) new evidence has made possible a more satisfactory interpretation of some puzzling features of the petrogenesis. An important problem raised in the paper was the origin and nature of the rock which was contact metamorphosed to give an olivine-bearing hornfels. At that time, from microscopic work, I could see no alternative to calling it a " basic argillite ". Apart from the difficulties arising from the fine grain-size of the rock, this idea developed by taking the rather over-simplified view that the Neranleigh-Fernvale Group consisted mainly of shales and greywackes coupled with the generally accepted view that basic volcanic rocks are absent from this Group except in the Brookfield area, some

twelve miles to the south. Hence I wrote, " In this area there are no known earlier basalts which could have formed the hornfels " (p. 382).

A few months ago the area was visited by Professor Allan F. Wilson in my company and he recognized criteria for deformed pillow lavas within 20 feet of the contact with the gabbro. Moreover, the pillows are associated in several places with the typical calcic inter-pillow "sludge". Laminated rocks are also present and in such a sequence they are now thought to be basic tuffaceous sediments. The whole assemblage of pillows, their calcic interpillow material and the tuffaceous sediments has suffered some deformation prior to the intrusion of the gabbro. Some pillows are drawn out into elongate ellipsoidal bodies and some of the calcic interpillow material has migrated and transgressed the metavolcanic sequence. It was this pre-gabbro metamorphism, together with the presence of the laminated rocks, which helped to make me think only of a sedimentary and not a basic igneous origin for the hornfels.

In view of the new field evidence, the mode of origin of the olivine-bearing hornfels becomes reasonably acceptable. It was formed by the contact metamorphism of a deformed basic lava, not of a " basic argillite ".

The chemical analyses published in the original paper now take on added significance. The analysis of the basic argillite (Phillips, 1959, Table 1, No. 3) is now considered to be that of portion of a metamorphosed basic pillow lava and the C.I.P.W. norm upholds this interpretation (see this letter, Table 1,

TABLE 1

	1	2
SiO_2	47·68	44·66
TiO_2	0·64	1·22
Al_2O_3	17·21	16·62
Fe_2O_3	4·52	4·02
FeO	6·64	7·18
MnO	0·11	0·14
MgO	7·11	12·35
CaO	12·94	10·36
Na_2O	1·74	1·72
K_2O	0·44	0·46
$H_2O + 110°$ C	0·89	1·15
$H_2O - 110°$ C	0·09	0·05
CO_2	Nil	Nil
P_2O_5	0·07	0·16
	100·08	100·09
Norms.		
or	2·8	2·8
ab	14·7	14·2
an	37·8	36·4
di wo	10·7	6·0
di en	6·9	4·3
di fs	3·0	1·2
hy en	10·2	5·7
hy of	4·4	1·5
ol fo	0·4	14·4
ol fa	0·2	4·3
mt	6·5	5·8
il	1·2	2·3
ap	0·3	0·3

1. Deformed basic lava from 150 yards west of the contact with gabbro, Cedar Creek at 917084 Samford Sheet. Earlier interpreted as the " basic argillite " (Phillips, 1959, Table 1, No. 3). Analyst—W. H. Herdsman.

2. Olivine-bearing hornfels at the contact with gabbro at 917084 Samford Sheet. Analyst—W. H. Herdsman.

No. 1). The analysis and C.I.P.W. norm of the " argillite " are similar to, but differ in certain respects from, the analysis and C.I.P.W. norm of the olivine-bearing hornfels (see Phillips, 1959, Table 1, No. 4, and this letter Table 1, No. 2). The composition differs, however, no more than one would expect a portion of a pillow (perhaps slightly contaminated with migrating interpillow calcic material) to differ from a more fortuitously selected sample of basic rock collected some 150 yards across the strike of a basic metavolcanic succession. Because of this, the idea that the contact metamorphism was practically isochemical is still the favoured hypothesis, and because of the new field evidence the relationship between hornfels and source rocks is a much simpler and more credible one.

Finally, I would like to say that at all times Associate Professor J. F. G. Wilkinson of the University of New England, thought from chemical evidence, without having seen the contact area, that the hornfels had been formed from some kind of basaltic rock, and I would like to thank him and Professor Wilson for the help they have given me.

EVAN R. PHILLIPS.

DEPARTMENT OF GEOLOGY,
UNIVERSITY OF QUEENSLAND,
ST. LUCIA,
BRISBANE.
6th June, 1961.

REFERENCE

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AN EARLY REFERENCE TO ISOSTASY

SIR,—The " Examination of Dr. Woodward's Account of the Deluge etc." is a rare pamphlet written by Dr. J. Arbuthnott and printed in London in 1697. The result of this " Examination " was very unfavourable to Dr. Woodward. In elegant language Arbuthnott, renowned wit, accomplished satirist and collaborator with Pope, Swift, and Gay, utterly demolished Woodward's edifice.

It will be remembered that, notwithstanding a denial of his obligation to postulate the origin of the waters of the Flood, Woodward did in fact claim that it came out of an internal cavity or abyss at the centre of the earth. It was this vacant hollow which excited Arbuthnott's interest. Nor is this surprising when we recall that this space had to remain empty until it was roofed by the reformed solid crust and to provide the receptacle into which the flood waters were to descend again when they had completed their business of temporarily transforming the earth into a thick suspension of earthy matter.

By a straightforward calculation based upon the reasonable assumption that there would be required at least a bulk of water equal to that of solid matter if any reasonable degree of fluidity of the mixture was to be achieved, Arbuthnott computed that the volume of water needed would envelop the earth to a depth of 450 miles. This quantity of water would occupy a sphere of 5,000 miles diameter with a centre coincident with that of the earth. Had he taken the modern estimate of the earth's radius in place of the then current figure of 3,600 miles his figures would have been increased by some 10 per cent.

The difficulties attendant upon the support of a liquid shell above an empty space naturally occurred to Arbuthnott, but what is more interesting is what he has to say about a solid carapace arching over such a vacuity. It is necessary, however, to grant him what he did not specifically claim, namely that his remarks would apply equally had this space been filled with a heavy fluid. Now, Woodward had interpreted horizontal planes of stratification as interfaces between concentric shells of materials of increasing density towards

the centre of the earth and formed by settlement of disaggregated particles from a suspension. Writing of bedding planes formed in this way, Arbuthnott proceeds:

"*I shall beg leave to add one reason why I think they would not be smooth a-top, and that is . . . to make them smooth* [i.e. planar] *each solid Column, or to speak more properly, each truncated Sector must have been of the same bulk and the Matter of the same density; of the same bulk, else there would have been outward or inward Inequalities, where there was a difference in Extension; and of the same density, else the several columns could not balance themselves exactly in the subsidence, but one would have been higher than the other; both then would have provided Mountains and Valleys. Now since there were antediluvian Mountains it would seem that the Columns of the Earth had not been made up of Matter with both the aforementioned Properties.*"

Thus Arbuthnott stated the conditions for the existence of a perfectly uniform spherical shell forming the outer skin of a uniform sphere containing a concentric perfect spherical cavity. He recognized that the surface of this inner sphere constituted what a modern geophysicist would call the surface of compensation and that mountains and valleys could occur above it if they formed the summits of columns of different density. This corresponds to Pratt's isostatic model. He also came near to Airy's conception of isostatic balance when he realized that sectors of equal density differing in bulk would result in "outward *or* inward Extensions". Had he replaced "or" by "and" the likeness to Airy's model would have been complete.

It is interesting to see how this penetrating critic thought along the same lines as the founders of the modern notions of isostasy, a century and a half before the basis of that fruitful conception was presented to geologists.

Department of Geology, Frank Hodson.
University of Southampton.
27th June, 1961.

THE MAAM FAULTS, CO. MAYO

Sir,—Professor J. G. C. Anderson remarks (Geol. Mag., 1960, xcvii, p. 511) that the north-westerly swing of the Wenlockian strike west of Cregganbaun is not "undoubtedly" related to the Maam Faults. The map (Pl. XIX) in a recently published paper on the Ordovician between Cregganbaun and Killary Harbour (Stanton, 1960) clearly shows them to be due to one and the same cause, the Maam Disturbance. As described in the paper, this is a north-westerly trending vertical shear zone at least 25 miles long with a dextral displacement of 1·7 to 3·5 miles. In the rigid Mweelrea Grits the Disturbance is represented by the Maam tear faults, but on passing into the incompetent slates and greywackes of the Glenummera Slates and Sheefry Grits the faults die out and are progressively replaced by a zone of flexure (c.f. the upper and lower boundaries of the Glenummera Slates). Here both strata and cleavage swing round to the north-west, the strata being attenuated and the regional cleavage being accentuated to the point of schistosity. The Wenlockian strata west of Cregganbaun cross the Maam Disturbance and are involved in the flexure; hence their change of strike. The displacement of the Wenlockian caused by the flexure is rather more than 2 miles, which is normal for the Maam Disturbance.

Professor Anderson suggests that: "the Maam fractures in fact ... developed very late and may in fact be due to post-Carboniferous N.-S. Armorican compression; the arcs, on the other hand, are pre-Carboniferous." Since the Maam Disturbance caused the simultaneous formation of both the Maam Faults and the Cregganbaun are this statement must be in error. In fact, the evidence from the Ordovician south of Cragganbaun indicates that the Maam Disturbance developed in step with the regional folding. To quote only two examples: there is no sign in the Sheefry Grits of distortion of the regional Caledonian cleavage such as would result from later large-scale

shearing ; and the many acid dykes in the flexure near Cregganbaun, although mostly well-cleaved, include a number of uncleaved dykes (Stanton, 1960, p. 289). Movement on the Maam Disturbance thus seems to have ceased before the final stages of emplacement of the Corvock Granite, from which the acid dykes derived.

There is strong evidence (Stanton, 1960, pp. 278, 282 ; Fig. 2) that preliminary movements on the line of the Maam Disturbance, between Ben Gorm, Fin Lough, and Ben Lugmore, took place during the deposition of the Mweelrea Grits. As the Disturbance is approached from the west both the First and Second Slate horizons thin out remarkably, but immediately the fault zone is crossed they reappear at maximum thickness, and an extra slate horizon is present overlying the welded tuff bed T.B.5. At this locality T.B.5. was involved in contemporaneous slumping, the only case of slumping affecting a Mweelrea Tuff Bed in the whole area. These facies changes were probably caused by downthrow to the north-east, repeated at least three times, along the line of the Disturbance. Far from being a late feature, therefore, the Maam Disturbance was already affecting sedimentation in the Middle Ordovician.

W. I. Stanton.

Servicos de Geologia e Minas,
Luanda,
Angola.
April, 1961.

REFERENCE

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Sir,—Dr. Stanton's letter provides an example of the kind of misinterpretations that can arise through considering structural events on a local rather than a regional basis.

The Maam Fault is only one of a system of transcurrent faults in South Mayo which are later than the regional schistosity and two phases of the regional folding, although a late cleavage may be locally produced along them. In fact, the Maam Fault itself, continued north-west, dextrally displaces the Highland Boundary fracture-zone and the Old Head Series (Old Red Sandstone) by about four miles which is of the same order as the movement recognized by Dr. Stanton further south.

That some of the acid dykes near Cregganbaun are not schistose is no proof that they are later than the fault. Where these dykes were unaffected by the regional metamorphism they would resist fault-cleavage just as have some of the quartzites of both Ordovician and Wenlock age. If the Cregganbaun flexure is caused mainly by the fault, it is curious that an equally marked bend south of Croagh Patrick should have developed without any such fault, and also that other faults of the same system should not have produced flexuring.

As the fault is transcurrent it is not surprising that there should be a change of facies across it. The possibility cannot be excluded (as with many other such faults) that the fault follows an ancient zone of disturbance but it cannot be accepted that the main movement was pre-Old Red and responsible for the bending of the post-Wenlock folds; the bending of the folds is due to cross-folding which is also identifiable in the minor structures.

J. G. C. Anderson.

Department of Geology,
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Cardiff.
14th August, 1961.

REVIEWS

The Relief and Drainage of Wales [1]: an Essay Review

By O. T. Jones

The High Plateau of Wales is the most striking feature in the relief of the country. In 1956 (1957) Dr. Brown put forward the view, now elaborated in his book, to the effect that its extension to the foot of the great escarpment which bounds the South Wales Coalfield in Breconshire and Carmarthenshire is part of the same surface as that of the South Wales Coalfield. However, nothing can be more certain than that the South Wales Coalfield area and the High Plateau are two surfaces of different origin. In 1948 I mentioned some geological evidence in favour of this opinion. Since it has escaped the author it may be useful, therefore, to restate that evidence more fully.

This great plateau, which surrounds the mountains of North Wales at a level of around 2,000 feet, slopes gently southwards through Central Wales toward the foot of the escarpment at the foot of the Brecon Beacons at a level of about 1,400 feet. This is probably the largest continuous escarpment in the British Isles. Apart from the mountains of North Wales, which rise conspicuously above the plateau level, there are some other summits which in any good view can be seen to stand out clearly above the surrounding plateau. Such are the Berwyn Mountains, Plynlimon, probably Radnor Forest and some smaller ones. The plateau can be seen in profile for miles from good view points, and apart from certain valleys which are deeply entrenched in it, is obviously a continuous feature, and it is impossible to regard it as anything but one surface from North Wales to the foot of the Beacons.

From their foot the ground rises sharply to the crest of the escarpment, composed of southerly dipping Upper Old Red Sandstone, and then descends in a series of long dip slopes and smaller escarpments to the Pennant Plateau in the middle of the South Wales Coalfield. Although the dip of the strata in the southern part of the South Wales Coalfield is reversed, the surface descends southwards from the Pennant Plateau to the crest of another escarpment in Old Red Sandstone, which defines the southern limb of the Coalfield basin. This escarpment is not so high as that of the Brecon Beacons but is, nevertheless, an impressive feature. From its foot the lower undulating area known as the Vale of Glamorgan extends to the coast of the Bristol Channel.

Mr. Brown maintains that the High Plateau of Wales and the surface of the South Wales Coalfield belong essentially to one surface, and since the general level of the South Wales Coalfield is not very different from that of the High Plateau in North Wales, there is therefore no evidence that the High Plateau has been sensibly tilted since it was formed, in the manner that I had suggested some years earlier. The point at issue is the significance of the escarpment that surrounds the coalfield in terms of the geological history of the region.

At the northern edge of the Vale of Glamorgan, Triassic breccias resting on Carboniferous rocks are tucked in at the foot of the escarpment, which rises several hundred feet above them. The escarpment was therefore in being when the Upper Trias was laid down, and it must be presumed that it was developed some time in the New Red Sandstone period between late-Carboniferous and late-Triassic times.

[1] *The Relief and Drainage of Wales. A Study in Geomorphological Development.* By Eric H. Brown. pp. xii + 186, 24 pls., 47 figs., and folded map. University of Wales Press. Cardiff, 1960. Price 30*s*.

During the New Red Sandstone, owing apparently to the climatic conditions of that period, there was extensive reddening of rocks that formed the surface during that period of erosion.

This reddening has been observed in many parts of the country ranging from South Wales and the Forest of Dean to the extreme north of Scotland. Some years ago I gave a list of places (1948) where this reddening had been observed, many of which I had myself observed. To these can now be added several localities in the Llanidloes district near the Van mine at an elevation of about 1,400 feet. This last area lies in fact in the direct continuation of the Cheshire Basin, but more than 30 miles distant from the nearest Triassic rocks.

As proof that this reddening is not due to infiltration of iron oxide down from the Trias as Strachan and later Sibly (1919) once thought, it has been found that the Coal Measures of the North-east of England, which are separated from the Trias by several hundred feet of Magnesian Limestone, are reddened in the same way as those of Lancashire, described by Trotter (1954).

From the edge of the Vale of Glamorgan the escarpment can be traced first eastward, then northward, always as a continuous feature except for some narrow valley breaches. The undulating low ground to the east of the escarpment extends towards the Forest of Dean where Trotter records much reddening of some of the older rocks, and attributes it to New Red Sandstone weathering. This area shows, therefore, the same kind of features as the pre-Triassic surface of the Vale of Glamorgan. The main difference is that it stands at a higher level.

Since both the eastern and the southern escarpments bear the same relation to areas which were eroded in the New Red Sandstone period, and since, moreover, both the escarpment and the lower grounds at its foot extend continuously from the eastern to the northern side of the coalfield, one is justified in assuming that the escarpment on the north side of the coalfield and the lower ground at its foot are of the same age and origin as the escarpment on the east and south. The level of the escarpment summit rises, however, from south to north, as does the general level of the lower ground, which the escarpment overlooks. The South Wales Coalfield and its surround to the edge of the escarpment form, therefore, a part of an ancient surface which was being eroded in New Red Sandstone times to form the lower ground which surrounds it on at least three sides. It is obviously, therefore, of different origin from the High Plateau, which extends without a break from North Wales to the foot of the Brecon Beacons escarpment. In fact the South Wales Coalfield area bears a similar relation to the plateau as the mountains of North Wales out of which that part of the plateau was also eroded.

It is probable that the present surface of the High Plateau is not identical with its condition in New Red Sandstone times, but it is unlikely that its relative elevations in North Wales and South Wales have undergone much change.

In my opinion my hypothesis that the High Plateau has been tilted since it was formed is the only one that successfully explains the main physical features of Wales and at the same time accounts for the amazingly close relation of present and former stream-directions to the present distribution of the Trias from Pembrokeshire to Cheshire. This relation led to the suggestion that the streams which enter Cardigan Bay did so because that Bay was probably a Triassic Basin like the Cheshire Basin. It also accounts, as no other hypothesis does, for the fact that the Prescelly mountains of Pembrokeshire rise at about 900 feet sharply out of a lower plain in the same way that the mountains of North Wales rise out of the High Plateau at a much higher elevation. Mr. Brown's view of the relief of Wales does not begin to afford an explanation for any of the above relationships.

One gathers from the dust cover that the study is " in the tradition of Henri Baulig ". Professor Baulig is an apostle of Eustatism (eg., 1935). In view of what is known of the present altitude of contemporary surfaces of Tertiary and Pleistocene ages this doctrine cannot possibly be applied to the

British Isles. Undue reliance on the Baulig tradition may, therefore, have led Mr. Brown to his peculiar views on the relief of Wales.

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SEDGWICK MUSEUM,
CAMBRIDGE.

ECOLOGY AND DISTRIBUTION OF RECENT FORAMINIFERA. By F. B. PHLEGER. viii + 297 pp., 83 figs., 11 pls. The John Hopkins Press, Baltimore. 1960. Price 60*s*.

This book provides a timely summary of modern work on the distribution of Recent foraminifera. It brings together a great deal of information otherwise only available from a widely-scattered literature. Benthonic, living benthonic, and planktonic faunas are dealt with separately ; there is a brief introductory chapter on principles and methods, and a short concluding discussion. The difficulties of applying the results so far obtained to the interpretation of fossil assemblages are clearly recognized. The literature of the last two decades is comprehensively covered by the list of references, which is complete up to and including part of 1960. There is a wealth of tables and diagrams, but in places these seriously disrupt the text.

The chief geological interest of this work lies in its relevance to the interpretation of post-Mesozoic, and particularly post-Palaeogene, foraminiferal faunas.

B. M. F.

AN INTRODUCTION TO THE SCIENTIFIC STUDY OF THE SOIL. By the late NORMAN M. COMBER. 4th Edn. (1960), revised by W. H. Townsend. Edward Arnold, Ltd., London. 232 pp. 18*s*. net.

Probably as a result of the long domination of pedology by agriculture, there is gross lack of reasonably priced texts which approach the subject from fundamental principles.

The new edition of the *Scientific Study of Soil* unfortunately does not fill this gap. It is intentionally aimed at the general agriculturalist and certainly, for his purposes, gives a useful and readable account of standard material. In addition there is a good chapter on the literature of the subject and hints on its use which will be of great value to the beginner. Furthermore, the book is well produced and sells at a price which compares very favourably with those of some familiar American texts covering similar ground.

With the exception of the chapter on the literature of the subject, however, the book has not a great deal to offer to the geologist who requires a short account of soils as soils rather than as media for crop growth. He will find only superficial treatment of those aspects of pedology in which he is most likely to be interested ; these include, for example, weathering processes, the colloid chemistry of silicate minerals, the movement of water and the development of lateritic soils.

R. M. S. P.

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